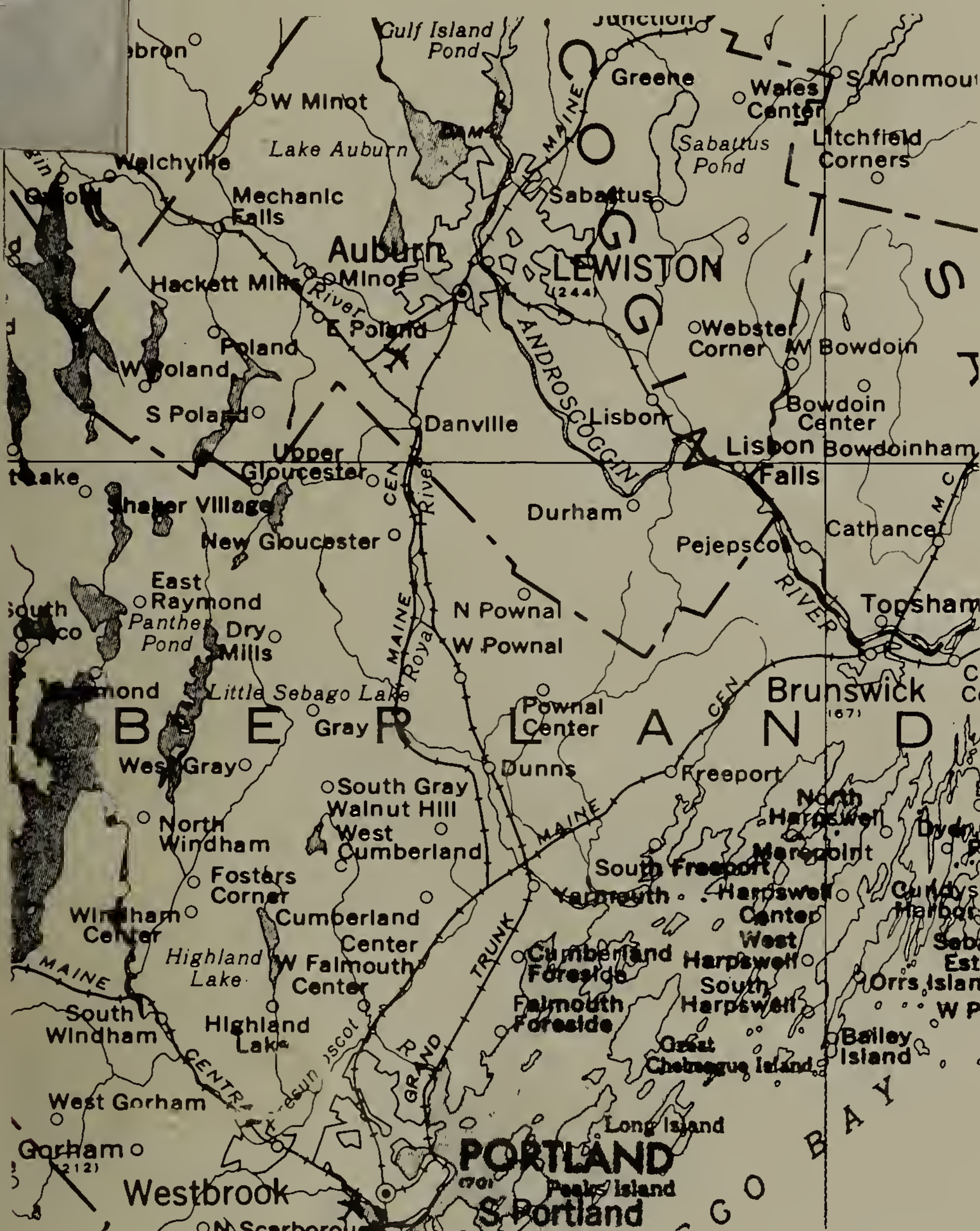

78th Annual Meeting
October 17-19, 1986

Bates College
Lewiston, Maine

Guidebook for Field Trips in Southwestern Maine

New England Intercollegiate Geological Conference

SAR System® SYNTHETIC APERTURE RADAR IMAGERY **Produced for the United States Geological Survey under USGS contract number 14-08-0001-21382 by Aero Service Division, Western Geophysical Company of America.**

Mosaic prepared from synthetic-aperture airborne radar image strips flown during September and/or October 1982. Flight height 37,000 feet above mean sea level. Depression angle range 11°-17°. Image mosaic controlled to identified points scaled from corresponding 1:250,000-scale map.

Mosaic compiled in February 1984

Information on indexing, cost, and availability of radar imagery may be obtained from USGS EROS Data Center, Sioux Falls, South Dakota 57198.

Produced for the United States Geological Survey under USGS contract 14-08-0001-21748.

Flown by INTERA Technologies, Inc., Houston, Texas, using the digital high resolution, synthetic aperture, STAR-1 radar system, operating from a flight altitude of 33,000 feet (9,145 meters) above mean sea level. Image strips, 46 kilometers wide, were flown on 20 kilometer line spacings. Separate "near range" and "far range" image mosaics were prepared for, and reconciled to, each of the 38, 1:250,000 - scale USGS topographic base maps in the project area. Photogrammetry and mosaic compilation performed by Simulation Systems, Inc., Division of MARS Associates, Inc., Phoenix, Arizona.

Information on indexing, cost, and availability of radar imagery may be obtained from USGS EROS Data Center, Sioux Falls, South Dakota 57198

Scale 1:250,000

5 0 5 10 15 20 Statute Miles

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Before visiting any of the sites described in the New England Intercollegiate Geological Conference guidebooks, you must obtain permission from the current landowners.

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New England Intercollegiate Geological Conference

78th Annual Meeting

Bates College
Lewiston, Maine

October 17, 18, and 19, 1986

GUIDEBOOK FOR FIELD TRIPS IN SOUTHWESTERN MAINE

Donald W. Newberg, Editor
Department of Geology
Bates College
Lewiston, Maine 04240

As long as supplies last extra copies of this guidebook are available at a cost of $12.00 which includes postage and handling. They may be obtained by writing:

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Foreword

A favorite passage of mine from Mark Twain's Life on the Mississippi reads as follows ....

> "In the space of one hundred and seventy-six years the Lower Mississippi has shortened itself two hundred and forty-two miles. That is an average of a trifle over one mile and a third per year. Therefore, any calm person, who is not blind or idiotic, can see that in the Old Oolitic Silurian Period, just a million years ago next November, the Lower Mississippi River was upward of one million three hundred thousand miles long, and stuck out over the Gulf of Mexico like a fishing rod. And by the same token any person can see that seven hundred and forty-two years from now the Lower Mississippi will be only a mile and three-quarters long, and Cairo and New Orleans will have joined their streets together and be plodding comfortably along under a single mayor and a mutual board of aldermen. There is something fascinating about science. One gets such wholesale returns of conjecture out of such a trifling investment of fact."

These words seem appropriate here. The winter conjecturing in which many of us delight is based upon an investment of summers' observational facts. And a purpose of the NEIGC over the years has been to gather people together to critically review and discuss these field observations. The trip guides which are included here therefore contain field data which the authors consider to have some particular importance. Diagrammatic and descriptive information is intended to help others locate these points of observation both during the 1986 Conference and in future time. Yet as Bob Tracy pointed out in last year's Guidebook ....

> "The fact that a locality is described in this guidebook does not imply that the public has access to the locality."

Anyone using the guides contained in this volume is encouraged to seek permission to visit the locations identified wherever it is appropriate to do so.

I would like to thank the various trip leaders who offered trips or were persuaded to lead one. Several pointed out the difficulty of trying to prepare a manuscript in the middle of the field season. I am particularly appreciative of the help given me by Theresa Shostak and Joyce Caron of Bates College in the preparation of the Guidebook. Arthur Griffiths of Twin City Printery made many helpful suggestions as well. Finally, Judith Marden, Director of Special Projects and Summer Programs at Bates, and Suzanne Shaw were very helpful in obtaining lodging information and in scheduling facilities to be used during the Conference. I thank them both.

Donald W. Newberg, Editor

Table of Contents

THE USE OF GEOPHYSICAL EQUIPMENT IN HYDROGEOLOGIC INVESTIGATIONS, AND THE MEASUREMENT OF STREAM DISCHARGE

Dorothy H. Tepper[1/], F.P. Haeni[2/], and Carole D. Johnson[1/]

Meeting place and time:	parking lot on the northeastern side of Merrill Gymnasium, Bates College, at 8:15 a.m., Friday, October 17. *Private vehicles will be needed for transportation to the Auburn gage house.

INTRODUCTION

The use of selected geophysical equipment in hydrogeologic investigations, and the measurement of stream discharge will be presented during this two-part technical session.

The geophysics session will involve field demonstrations at Bates College by Survey (U.S. Geological Survey), MGS (Maine Geological Survey), and MDEP (Maine Department of Environmental Protection) personnel. The following geophysical techniques and equipment will be demonstrated: seismic refraction (1-channel and 12-channel seismographs); ground-penetrating radar; direct-current resistivity; and electromagnetics (terrain conductivity and resistivity). The field trip group will be split into smaller groups that will spend approximately 1 hour at each of the concurrent demonstrations of the above equipment. Principles, hydrogeologic uses, limitations, interferences, field setup, and data interpretation for each of the geophysical techniques will be discussed.

The stream-discharge-measurement session will be run concurrently with the geophysics session. It will involve a 1 1/2-hour demonstration of discharge measurements at the Survey's gaging station on the Androscoggin River at Auburn. There will be a discussion of the Survey's stream-gaging network in Maine, an explanation of the equipment in the gaging station, and a demonstration of a cable-car discharge measurement.

The following station descriptions provide summaries of information presented at the geophysics and stream-discharge measurement sessions. In addition, a list of selected references on geophysical methods and the use of integrated geophysical techniques in hydrogeologic investigations is presented at the back of this article.

1/ Hydrologist, U.S. Geological Survey, Water Resources Division, Augusta, Maine.

2/ Hydrologist, U.S. Geological Survey, Water Resources Division, Hartford, Connecticut.

STATION DESCRIPTIONS

Station #1: Seismic-Refraction Techniques and Equipment

Multi-Channel Seismic Refraction

Demonstration by: Dorothy H. Tepper, Hydrologist
U.S. Geological Survey
Augusta, Maine

Physical property measured: The seismic-refraction method measures the time it takes a compressional sound wave to travel through the layers of the earth to detectors (geophones) placed on the land surface. The subsurface geology can be interpreted by measuring the traveltime of the sound wave and then applying the laws of physics that govern the propagation of sound through layered media.

Hydrogeologic uses: Seismic-refraction techniques can be used to determine:

- depth to water table in unconsolidated material,
- depth to and configuration of bedrock surface beneath unconsolidated material,
- depth to crystalline rocks beneath sedimentary rocks, and
- saturated thickness of aquifer material.

Limitations:

- The velocity of each successive layer must increase with depth to detect it with seismic refraction techniques.
- Layer velocities must be sufficiently different to distinguish individual layers.
- Thin layers may not be detectable even if the velocity constraints listed above are met.
- Long spreads or large distances from the shot point to the first geophone may be required, depending on the depths to the subsurface layers of interest.
- Explosives may be needed in order to obtain high-quality record.
- Different combinations of subsurface structure or layering can result in similiar time-distance plots. Because the solution is not unique, information on the hydrogeology in the area is helpful for calibration. If calibration data are unavailable, more than two shots could be fired on the line to increase data redundancy, thereby increasing the reliability of the data.
- A high-velocity layer at the land surface, such as frozen ground, will not allow distinction of layers of lower velocity beneath it. This technique, therefore, has limited applications in permafrost zones.
- Depending on the particular seismograph used, there may be no permanent record of the output (wave forms).

Interferences: Interference problems, resulting in poor-quality record, can be caused by:

- motion of nearby vehicles or heavy machinery,
- wind and associated tree-root movement,

- high humidity (can cause increased problems with electrical interferences), and
- nearby powerlines or other sources of electromagnetic fields.

Approximate cost of field equipment:

- A state-of-the-art 12-channel, signal-enhancement seismograph and accessories cost approximately $10,000 to $30,000.
- Costs for sound sources differ greatly depending on the type of source used. For example, if explosives are used, a drill may be required for making the shot holes. Training in the safe handling of explosives should be provided for personnel. Depending on the type of explosives used, the cost per shot may range from approximately $5 to $15.

Field crew required: A minimum of two people is required, but a crew of three people is preferable.

Estimated daily production:

Field:
- In an open area with deep valleys, approximately 0.5 to 0.75 miles of seismic data can be collected, using overlapping 1,100-foot spreads and multiple shot points.
- In a wooded area with shallow valleys, approximately 0.25 to 0.5 miles of seismic data can be collected using overlapping spreads and multiple shot points.

Office:
- Approximately 1 day of interpretation time should be planned for each day of field work.

Data interpretation:

- An inverse modeling program (Scott and others, 1972), which is based on the delay time method and a ray-tracing modeling technique, is commonly used. Output includes a time-distance plot, apparent velocities for each layer, depths to each layer beneath each shot point and geophone, and a subsurface profile.
- Numerous other interpretation programs (see Ballantyne and others, 1981) which use various methods and modeling techniques are available for use with hand-held calculators and microcomputers, minicomputers, and mainframes.

Selected references:

Ballantyne, E.J., D.L. Campbell, S.H. Mentemeier, and Ralphe Wiggins, (eds.), 1981, Manual of geophysical hand-calculator programs, vol. 2: Society of Exploration Geophysicists, Tulsa, Oklahoma.

Birch, F.S., 1976, A seismic ground-water survey in New Hampshire: Ground Water, v. 14, no. 2, p. 94-100.

Haeni, F.P., 1978, Computer modeling of the ground-water availability of the Pootatuck River Valley, Newtown, Connecticut: U.S. Geol. Surv. Water Resources Investigations Report 78-77, 64 p.

_____, 1986, Application of seismic refraction methods in ground-water modeling studies in New England: Geophysics, v. 51, no. 2, p. 236-249.

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Mooney, H.M., 1980, Handbook of engineering physics, volume 1: seismic: Bison Instruments, Inc., Minneapolis, Minn., 193 p.

Morrissey, D.J., 1983, Hydrology of the Little Androscoggin River valley aquifer, Oxford County, Maine: U.S. Geol. Surv. Water-Resources Investigations Report 83-4018, 79 p.

Pakiser, L.C., and R.A. Black, 1957, Exploring for ancient channels with the refraction seismograph: Geophysics, v. 22, no. 1, p. 32-47.

Reynolds, R.J., and G.A. Brown, 1984, Hydrogeologic appraisal of a stratified-drift aquifer near Smyrna, Chenango County, New York: U.S. Geol. Surv. Water-Resources Investigations Report 84-4029, 53 p.

Scott, J.H., 1973, Seismic refraction modeling by computer: Geophysics, v. 38, no. 2, p. 271-284.

Scott, J.H., B.L. Tibbetts, and R.G. Burdick, 1972, Computer analysis of seismic-refraction data: U.S. Dept. of Interior, Bureau of Mines Report of Investigations RI 7595, 95 p.

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Tepper, D.H., J.S. Williams, A.L. Tolman, and G.C. Prescott, Jr., 1985, Hydrogeology of significant sand and gravel aquifers in parts of Androscoggin, Cumberland, Franklin, Kennebec, Lincoln, Oxford, Sagadahoc, and Somerset Counties, Maine: Sand and gravel aquifer maps 10, 11, 16, 17, and 32: Maine Geol. Surv. Open-File Report 85-82a, 106 p.

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Single-Channel Seismic Refraction

Demonstration by: Craig Neil, Research and Planning Associate
Maine Geological Survey
Augusta, Maine

Physical property measured: same as multichannel seismic refraction

Hydrogeologic uses: same as multichannel seismic refraction

Limitations: In addition to the limitations listed for multichannel seismic refraction, the following are limitations of the single-channel method:

- Because sound sources typically used for single-channel work are not high-energy and therefore do not generate strong signals, this technique generally works best where the depths to the layers of interest are within 50 to 100 feet of the land surface.

- Each spread is typically only 200 to 300 feet long, so multiple spreads will be required to profile a large area.
- Depending on the particular seismograph used, there may be no permanent record of the wave form.

Interferences: same as multichannel seismic refraction

Approximate cost of field equipment: A state-of-the-art signal enhancement single-channel seismograph with accessories costs approximately $4,500 to $5,500.

Field crew required: Two people are required.

Estimated daily production:

Field:
- In a relatively open area, approximately 10 to 15 spreads can be run (this should allow some time for preliminary data interpretation).

Office:
- Depending on the hydrogeologic complexity, each spread will take approximately 1 to 3 hours to interpret.

Data interpretation: Many programs are available for hand-held computers and micro-computers. The program that is commonly used by both the Maine Geological Survey and the Maine Department of Environmental Protection was written by Mooney (1980). Output includes calculated velocity for each layer and depth to each layer under the two geophones.

Selected references:
In addition to those listed under multichannel seismic refraction:

Heeley, R.W., and B.A. Marshall, 1985, The use of geophysical techniques in an accelerated search for ground water in the Connecticut River valley, Massachusetts: in Nielson, D.M. and M. Curl (eds.), Surface and borehole geophysical methods in ground water investigations-second national conference and exposition: National Water Well Association, Worthington, Ohio, 424 p.

Sverdrup, K.A., 1986, Shallow seismic refraction survey of near surface ground water flow: Ground Water Monitoring, v. 6, no. 1, p. 80-83.

Station #2: Ground-Penetrating Radar Techniques and Equipment

Demonstration by: David G. Johnson, Hydrologist
U.S. Geological Survey
Boston, Massachusetts

The following discussion is based largely on information from Benson and others (1982).

Physical property utilized: Radar waves are reflected from interfaces between materials having sufficiently different dielectrical properties. A continuous cross-sectional profile of shallow subsurface conditions can be produced based on variations in the return signal.

Hydrogeologic uses: Radar techniques can be used to determine the following:

- subsurface structure and stratigraphic changes
- moisture content of subsurface materials
- depth to the water table
- discontinuous clays at depth
- buried stream channels
- buried waste materials
- buried utilities
- depth to the bedrock surface
- bedrock fractures

Limitations:

- The principal limitation of radar is the depth of signal penetration, which is determined primarily from the attenuation produced from the sum of geometric scattering losses, electrical conductivity, and dielectric relaxation. Signal penetration is poor in conductive material and good in resistive material. Radar signal penetration capability is highly site-specific and can range from less than 3 feet to over 100 feet.
- Depending on the antenna (frequency) used, the resolution on the record may range from inches to several feet. High-frequency antennas (500 to 900 MHz) only provide shallow signal penetration but provide resolution of features on the scale of a few inches. In contrast, low-frequency antennas (80 to 125 MHz) can provide better signal penetration but can only provide resolution of features on the scale of a few feet or larger.
- Depth is not measured directly. It is calculated based on the velocity of radar waves in various materials and on travel time back and forth to the reflector.
- Depth calibration has to be done carefully. If conditions change, the depth calibration will be affected. In addition, the depth scale is usually nonlinear.

<u>Interferences</u>: Interference problems, resulting in poor-quality record, can be caused by:

- system noise: improper cable placement, locating antenna too close to towing vehicle
- overhead radar reflections: power lines, trees, buildings, etc. can affect lower frequency antennas that are not shielded on their top surfaces
- noise from surface factors: pieces of metal on the ground, topographic variations
- noise from subsurface features or buried debris
- external electromagnetic noise: nearby radio transmitters

<u>Approximate cost of field equipment</u>: A state-of-the-art ground-penetrating radar system and accessories cost approximately $17,000 to $50,000.

<u>Field crew required</u>: Depending on whether the antenna is towed by a vehicle or pulled by hand, two or three people will be needed. Experienced personnel are required due to the sophistication of the instrument and the technique.

<u>Estimated daily production</u>:

Field:
- For reconnaissance-level surveys, the antenna can be towed by a vehicle and data can be acquired at a rate of approximately 3 to 5 miles per hour. If more detailed surveys are required, the antenna can be hand-towed and data can be collected at a rate of approximately 0.3 to 0.5 miles per hour.

Office:
- See following discussion on "Data interpretation".

<u>Data interpretation</u>:

- Radar data are relatively straight-forward to interpret if hydrogeologic conditions are not complex and if there is a strong dielectric contrast between the features of interest and the surrounding material. As conditions become more complex, data interpretation becomes increasingly difficult and computer processing may be required.
- Graphical results can be printed in the field, allowing rapid qualitative and semi-quantitative analyses of the data, but experienced personnel are required for accurate interpretation. Radar data can be recorded on magnetic tape or other media which provides a back-up copy of the data, permits optimization of data quality, and can provide signal input to a computer or the control unit for various processing options. For example, analog- and digital-filtering techniques may be used to remove background or system noise. However, processing of data can be costly and may result in elimination of some important data.

Selected references:

Benson, R.C., R.A. Glaccum, and M.R. Noel, 1982, Geophysical Techniques for sensing buried wastes and waste migration: Environmental Monitoring Systems Laboratory, U.S. Environmental Protection Agency, Las Vegas, Nevada, p. 38-62.

Benson, R.C., and R.A. Glaccum, 1979, Radar surveys for geotechnical site assessment: in Geophysical methods in geotechnical engineering, specialty session, Amer. Soc. Civil Engineers, Atlanta, Georgia, p. 161-178.

Houck, R.T., 1984, Measuring moisture content profiles using ground-probing radar: in Nielson, D.M. and M. Curl (eds.), NWWA/EPA conference on surface and borehole geophysical methods in ground water investigations: Natl. Water Well Assoc., Worthington, Ohio, p. 637-653.

Olhoeft, G.R., 1984, Applications and limitations of ground penetrating radar: in Expanded abstracts, 54th annual meeting, Soc. Expl. Geophysicists, Atlanta, Georgia, p. 147-148.

Ulriksen, P.F., 1982, Application of impulse radar to civil engineering: Lund University of Technology, Lund, Sweden, 179 p.

Underwood, J.E., and J.W. Eales, 1984, Detecting a buried crystalline waste mass with ground-penetrating radar: in Nielson, D.M., and M. Curl (eds.), NWWA/EPA conference on surface and borehole geophysical methods in ground water investigations: Natl. Water Well Assoc., Worthington, Ohio, p. 654-665.

Wright, D.L., G.R. Olhoeft, and R.D. Watts, 1984, Ground-penetrating radar studies on Cape Cod: in Nielson, D.M. and M. Curl (eds.), NWWA/EPA conference on surface and borehole geophysical methods in ground water investigations: Natl. Water Well Assoc., Worthington, Ohio, p. 666-680.

Station #3: Direct Current Resistivity

Demonstration By: Andrews Tolman, Hydrogeologist
Maine Geological Survey
Augusta, Maine

Physical property measured: This technique measures the electrical resistivity of earth materials and the fluids in them.

Hydrogeologic uses: This technique is used to distinguish materials that differ significantly in electrical resistivity. Surveying can be done both horizontally (profiling) and vertically (sounding).
Example uses are to:

- determine the depth to the freshwater-saline water interface,
- determine the depth and thickness of significant clay layers,
- locate coarse, permeable material surrounded by fine silt or clay sediments, and
- locate contamination plumes.

Limitations:

- The solution may not be unique.
- It is difficult to differentiate between highly conductive materials.
- As the geology becomes more complex, the solution becomes more difficult.
- The electrodes must be driven into the ground. DC resistivity cannot be done on concrete.
- The method requires adequate space (the length of the line must be 3 to 12 times the depth of interest, depending on which configuration is used).
- Interpretation techniques assume a model of an infinite, horizontally-layered earth.
- There must be good electrical contact between the electrode and the ground. The method does not work effectively on dry or frozen ground.

Interferences: Stray currents, electromagnetic fields and conductive material can cause noise. Interferences include:

- aerial or buried power wires;
- water lines, pipes, railroad tracks, or metal fences; and
- inhomogeneities in shallow subsurface material.

Approximate cost of field equipment: $8,500 (signal-enhancement equipment)

Field crew required: This method can be done with a minimum of two people but three people are preferred.

Estimated daily production:

Field: • A field crew can complete 6 to 8 shallow soundings per day, or 15 profiles per day.

Office: • The time required for data interpretation varies depending on which method of interpretation is used.

<u>Data interpretation</u>: The apparent resistivity can be determined from the applied current and the measured voltage. These calculations can be done with a hand-held calculator.

- Values of apparent resistivity calculated in a profiling survey are plotted against distance. These data can be used to locate contamination plumes or variations in the stratigraphy.
- Values for apparent resistivity calculated in a sounding survey are plotted against the electrode ("A") spacing on log-log paper. Analysis of the plotted data requires a good understanding of the hydrogeologic conditions and of curve matching and/or the use of computer models.

<u>Selected References</u>:

Bisdorf, R.J. and A.A.R. Zohdy, 1979, Geoelectrical investigations with Schlumberger soundings near Venice, Parrish, and Homosassa, Florida: U.S. Geological Survey Open-File Report 79-841, 114 p.

Heigold, P.C., R.H. Gilkeson, K. Cartwright, and P.C. Reed, 1979, Aquifer transmissivity from surfical electrical methods: Ground Water, v. 17, no. 4, p. 338-345.

Kosinski, W.K. and W.E. Kelly, 1981, Geoelectric soundings for predicting aquifer properties: Ground Water, v. 19, no. 2, p. 163-171.

Mooney, H.M., 1983, Handbook of engineering geophysics; volume 2: electrical resistivity: Bison Instruments, Inc., Minneapolis, Minn., 193 p.

Stewart, M., M. Layton, and T. Lizanec, 1983, Application of resistivity surveys to regional hydrogeologic reconnaissance, Ground Water, v. 21, no. 1, p. 42-48.

_____, 1981, Electrical resistivity/hydraulic conductivity relations in glacial outwash aquifers: Water Resources Research, v. 17, no. 5, p. 1401-1408.

Urish, D.W., 1983, The practical application of surface electrical resistivity to detection of ground-water pollution: Ground Water, v. 21, no. 2, p. 144-152.

Van Nostrand, G., and K.L. Cook, 1966, Interpretation of resistivity data: U.S. Geol. Surv. Professional Paper 499, 310 p.

Zohdy, A.A.R., 1973, A computer program for the automatic interpretation of Schlumberger sounding curves over horizontally stratified media: National Technical Information Service, PB-232 703, Springfield, Va., 25 p.

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Zohdy, A.A.R., and R.J. Bisdorf, 1975, Computer programs for the forward calculation and automatic inversion of Wenner sounding curves: Springfield, Va., National Technical Information Service, PB-247, 265/AS p. 1-8.

Zohdy, A.A.R., G.P. Eaton, and D.R. Maybe, 1974 Application of surface geophysics to ground-water investigations: U.S. Geol. Surv. Techniques of Water-Resources Investigations, Book 2, Chapter D1, 116 p.

Station #4: Electromagnetic Methods

Terrain Conductivity

Demonstration by: William Aldrich, Geologist
Maine Department of Environmental Protection
Augusta, Maine

Physical property measured: This method measures the conductivity of earth materials and the fluids within them.

Hydrogeologic uses: The EM (electromagnetic) methods have the same applications as resistivity methods. The EM methods can be done on dry or frozen ground and do not require as much space as resistivity methods.

Limitations: The EM data cannot be interpreted if the problem involves more than 2 to 3 layers or involves complicated stratigraphy.

- The sampling depth is approximately equal to 1.5 times the coil spacing although it is also dependent on the coil orientation (vertical or horizontal), and the types of subsurface materials. Instruments are commerically available with coil spacings of 3.3, 13.1, 32.8, 65.6, and 131.2 feet.
- Because the conductivity reading is a bulk measurement of subsurface conditions down to the sampling depth, it has less resolution capability than direct current resistivity.
- In areas of very high conductivity, the measurements become nonlinear.

Interferences: Stray currents, electromagnetic fields, and conductive material can cause noise. Interferences include:

- aerial or buried power wires;
- water lines, pipes, railroad tracks, or metal fences; and
- atmospheric conditions (lightning).

Approximate cost of field equipment: There are a variety of terrain conductivity meters designed for different depths of penetration. The equipment costs range from $4,200 to $15,000, depending on the EM system. Features include:

- continuous recording instruments,
- hand-held or vehicle-mounted systems, and
- operation by one or two people.

NOTE: The remainder of this discussion refers to the Geonics EM34-3 terrain conductivity meter.[1]

Field crew required: This method requires two people.

Estimated daily production:

Field:
- In open areas, a field crew can complete one mile of profiling per day, with a station spacing of 100 feet.

Office:
- See discussion on "Data interpretation".

Data interpretation: Data can be interpreted manually or by computer. There are software (including graphics) packages commercially available for some microcomputers.

- Profiling data are plotted as the apparent conductivity at specific sampling depths as a function of distance.
- Sounding data are presented in a vertical electric section, which is then correlated to geohydrologic sections.
- Both types of data can be plotted and contoured using linear or logarithmic units. The apparent conductivity values measured in the field can be plotted using linear units. To plot the data in logarithmic units, the data must be converted to decibel units.

Selected references:

Barlow, P.M. and B.J. Ryan, 1985, An electromagnetic method for delineating ground-water contamination, Wood River Junction, Rhode Island: in Subitzkey, S. (ed.), Selected papers in the hydrologic sciences, U.S. Geol. Surv. Water-Supply Paper 2270, p. 35-50.

Duran, P.B., 1984, The effects of cultural and natural interference on electromagnetic conductivity data: in Nielson, D.M. and M. Curl (eds.), NWWA/EPA conference on surface and borehole geophysical methods in ground-water investigations: Natl. Water Well Assoc., Worthington, Ohio, p. 509-530.

1/ Use of trade names in this paper is for descriptive purposes only and does not constitute endorsement by the U.S. Geological Survey, the Maine Geological Survey, or the Maine Department of Environmental Protection.

Duran, P.B. and F.P. Haeni, 1982, The use of electromagnetic conductivity techniques in the delineation of ground-water contamination plumes, in The impact of waste storage and disposal on ground-water resources (Proceedings of the Northeast Conference, Ithaca, N.Y., June 28-July 1, 1982): U.S. Geol. Surv. and Center for Environmental Research, Cornell Univ., p. 8.4.1.-8.4.33.

Grady, S.L. and F.P. Haeni, 1984, Application of electromagnetic techniques on determining distribution and extent of ground-water contamination at a sanitary landfill, Farmington, Ct: in Nielson, D.M. and M. Curl (eds.), NWWA/EPA conference on surface and borehole geophysical methods in ground-water investigations: Natl. Water Well Assoc., Worthington, Ohio, p. 338-367.

Grantham, D.G., F.P. Haeni, and Karl Ellefsen, Forward modeling program for the inductive electromagnetic ground conductivity method: EM34.FOR: U.S. Geol. Surv. Open-File Report (in press).

Greenhouse, J.P. and D.D. Slaine, 1983, The use of reconnaissance electromagnetic methods to map contaminant migration: Ground Water Monitoring Review, v. 3, no. 2, p. 47-59.

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Very Low Frequency (VLF) Radio Wave Resistivity

Demonstration by: Carole D. Johnson, Hydrologist
U.S. Geological Survey
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Physical property measured: Resistivity of earth materials and the fluids in them.

Hydrogeologic uses: In addition to the uses listed for terrain conductivity:

- determination of layering vs. uniform conditions
- determination of layer thickness

Limitations: same as terrain conductivity

Interferences: In addition to the information listed for terrain conductivity, there can be noise associated with the VLF sound source. The survey site should be greater than 500 miles from the transmitter (naval communication centers).

Approximate cost of field equipment: This instrument costs approximately $8,000.

Field crew required: One person is required although two may work more efficiently.

Estimated daily production: same as terrain conductivity

Data interpretation:

- The VLF data are plotted as horizontal profiles. Plots of resistivity versus distance and phase angle versus distance are used to locate areas of low resistivity and to identify a layered earth.
- The data can be plotted in the same format as terrain conductivity data.

Selected references: In addition to the references listed for terrain conductivity:

Grantham, D.G., F.P. Haeni, and D.L. Mazzaferro, Forward modeling computer program for the very low frequency radio wave earth resistivity method: VLF.BAS: U.S. Geol. Surv. Open-File Report (in press).

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Station #5: Survey Stream-Gaging Station on the Androscoggin River at Auburn, Maine

Demonstration by: Derrill Cowing and Bill Bartlett, Hydrologists
U.S. Geological Survey
Augusta, Maine

Streamflow supplies water for domestic, commercial, industrial, and recreational uses. Streamflow records provide information on the availability of surface water and its variation in time, and the basic data required to develop reliable surface-water supplies. Gaging-station records from flood events are used as the basis for the design of bridges, dams, culverts, reservoirs for flood control, flood-warning systems, and are useful in the delineation of flood plains.

In the 1984 water year (October 1983 through September 1984), the Maine office of the Survey, in cooperation with other Federal and State agencies and with private industries, collected discharge records at 53 gaging stations, stage-only records at two gaging stations, and storage data for 17 lakes and reservoirs. This information, along with additional data on water quality at 11 gaging stations, water levels for 17 observation wells, and miscellaneous measurements, was published in an annual data report (Haskell and others, 1985). The following explanation of stage and water-discarge records and the data for the Auburn gage are excerpted from that report.

<u>Records of Stage and Water Discharge</u>

Records of discharge are obtained using a continuous stage-recording device through which either instantaneous or daily mean discharges may be computed for any time, or any period of time, during the period of record.

The data obtained at a gaging station on a stream consist of a continuous record of stage, individual measurements of discharge throughout a range of stages, and notations regarding factors that may affect the relationships between stage and discharge. These data, together with supplemental information, such as weather records, are used to compute daily discharges.

Continuous records of stage are obtained with analog recorders that trace continuous graphs of stage or with digital recorders that punch stage values on paper tapes at selected time intervals. Measurements of discharge are made with current meters using methods adopted by the Survey as a result of experience accumulated since 1880. These methods are described in standard textbooks, and in Survey publications by Rantz and others (1982) and Carter and Davidian (1968).

In computing discharge records, results of individual measurements are plotted against the corresponding stages, and stage-discharge relation curves are then constructed. From these curves, rating tables indicating the discharge for any stage within the range of the measurements are prepared. If it is necessary to define extremes of discharge outside the range of the current-meter measurements, the curves are extended using: (1) logarithmic plotting; (2) velocity-area studies; (3) results of indirect measurements of peak discharge, such as slope-area or contracted-opening measurements, and computations of flow over dams or weirs; or (4) step-backwater techniques.

Daily mean discharges are computed using the stages (gage heights) and the rating tables. If the stage-discharge relation is subject to change because of frequent or continual change in the physical features that form the control, the daily mean discharge is determined by the shifting-control method, in which correction factors based on the individual discharge measurements and notes of the personnel making the measurements are applied to the gage heights before the discharges are determined from the curves or tables. This shifting-control method also is used if the stage-discharge relation is changed temporarily because of aquatic growth or debris on the control. For some stations, formation of ice in the winter may so obscure the stage-discharge relations that daily mean discharges must be estimated from other information such as temperature and precipitation records, notes of observations, and records for other stations in the same or nearby basins for comparable periods.

At some stream-gaging stations, the stage-discharge relation is affected by the backwater from reservoirs, tributary streams, or other sources. This necessitates the use of the slope method in which the slope or fall in a reach of the stream is another factor in computing discharge. The slope or fall is obtained by means of an auxiliary gage set at some distance from the base gage.

For some gaging stations, there are periods when no gage-height record is obtained, or the recorded gage height is so faulty that it cannot be used to compute daily discharge. This happens when the recorder stops or otherwise fails to operate properly, intakes are plugged, the float is frozen in the well, or for various other reasons. For such periods, the daily discharges are estimated from the recorded range in stage, previous or following record, discharge measurements, weather records, and comparison with other station records from the same or nearby basins.

Discharge Record at the Auburn Gage

The record for the 1984 water year from the Auburn gage (table 1) includes information on station location, basin characteristics, regulatory control features, operation of the station, and extremes of data for the period of record. Daily discharge values and monthly and annual flow statistics are presented. Additional information is available from the Survey concerning flood frequency, low-flow frequency, and monthly and annual flow statistics.

Driving directions to gage: (transportation will be by private vehicles)

From Bates College campus, follow Russell Street west to the intersection of Route 202 (North Main Street). At the traffic light at this intersection, turn left (south) onto Route 202. Follow Route 202 through Lewiston. Just after crossing the North Bridge over the Androscoggin River, turn left on Route 136 (South Main Street). The gage house, which is a 5-foot by 5-foot concrete building, is located approximately 2 miles south of this intersection, on the east (left) side of Route 136.

TABLE 1. STREAMFLOW RECORD FOR THE 1984 WATER YEAR FOR THE ANDROSCOGGIN RIVER NEAR AUBURN, MAINE

LOCATION:--Lat 44 04'20", long 70 12'31", Androscoggin County, Hydrologic Unit 01040002, on right bank 1.5 miles downstream from Little Androscoggin River and 2.1 miles downstream from north bridge bewteen Auburn and Lewiston.

DRAINAGE AREA.--3,263 square miles

PERIOD OF RECORD.--Discharge: October 1928 to current year. Monthly discharge only for October 1928, published in Water-Supply Paper 1301.
Water-quality records: Water years 1966-75.

REVISED RECORDS.--WSP 781: 1930, 1933-34. WSP 1301: 1032-36: WDR ME-81-1: Drainage area.

GAGE.--Water-stage recorder. Datum of gage is 109.18 feet (National Geodetic Vertical Datum of 1929).

REMARKS.--Records good. Considerable diurnal fluctuation and some regulation by powerplants above station. Flow regulated by Rangeley, Mooselookmeguntic, Richardson, Aziscohos, Umbagog, Auburn, and Thompson Lakes and Gulf Island Pond (Reservoirs in Androscoggin River Basin) with major regulation at Errol Dam 136 miles upstream. Several observations of water temperature and specific conductance were made during the year.

AVERAGE DISCHARGE.--56 years, 6,192 ft /s (cubic feet per second).

EXTREMES FOR PERIOD OF RECORD.--Maximum discharge, 135,000 ft /s Mar. 20, 1936, gage height, 27.57 feet from rating curve extended above 76,000 ft /s on basis of slope-area measurement of peak flow and computation of flow over dam; minimum daily, 340 ft /s Sept. 28, 1941.

EXTREMES FOR CURRENT YEAR.--Maximum discharge, 62,500 ft /s Apr. 7, gage height, 17.27 feet; minimum

DISCHARGE, IN CUBIC FEET PER SECOND, WATER YEAR OCTOBER 1983 TO SEPTEMBER 1984
MEAN VALUES

DAY	OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP
1	2310	3160	13400	6380	4640	8570	8240	15300	48500	8330	3880	2200
2	1940	3070	10900	6050	4330	7950	9300	15900	44800	6750	3820	1990
3	2560	3060	9340	5810	4030	7510	9780	13100	39500	6640	3740	1980
4	2510	3410	7980	5550	4130	6320	10800	12400	36800	6830	2660	2790
5	2480	7160	6470	5650	5050	6080	13700	17700	30900	4070	2240	2880
6	2790	19000	7120	5600	5430	5900	45400	17200	24300	2580	2910	2770
7	3110	14700	12700	5850	5250	5960	59600	14600	20200	4920	3000	2760
8	2540	10200	14200	5740	5140	5500	43600	13600	17100	7650	2990	1960
9	2580	7980	10500	5410	4950	5490	33700	13300	14100	9580	2850	1760
10	3360	6680	9150	5190	4660	5470	26600	14300	10900	7610	2770	2570
11	3220	7580	8210	5110	4570	5000	21400	11600	9670	7010	2120	2830
12	2670	9490	6860	4960	4560	4640	17900	10800	8510	7140	1940	3250
13	3590	11400	9900	4850	4730	4940	18100	14000	7340	6670	2640	3150
14	3690	8880	29100	4620	5300	4780	19400	16100	7590	6500	2780	2970
15	3850	7170	38200	4770	5620	5160	20700	16000	6830	6080	2810	1970
16	3910	7450	28000	4730	7490	5410	20200	15200	7150	4510	2860	1160
17	4160	9910	20600	4710	9530	5620	23600	13800	6930	5390	3260	2760
18	3740	13400	16500	4630	10300	5500	27700	12000	6350	5440	2270	2850
19	3370	10200	14200	4600	9040	5770	27900	10800	6860	5550	1980	2800
20	3200	7770	10000	4840	9120	8230	25600	9160	6860	5920	2860	2840
21	2700	8370	7410	4730	8890	14900	22600	8240	6770	4760	2800	2910
22	1970	11800	7820	4620	8490	16200	20300	7560	6030	4370	2780	2420
23	769	12600	7230	4630	7660	16500	17100	7160	4260	4900	2830	2180
24	2950	10300	7070	4370	7870	15700	17000	7750	1020	4490	2930	2710
25	3120	15700	6640	4250	8860	13100	22100	9710	4980	3930	2000	2780
26	3020	29400	6700	4540	13300	11700	23600	7990	8300	4720	2300	2800
27	3310	27600	6590	4430	13700	10000	19600	7350	10800	3770	2880	2770
28	3250	15800	6560	4560	12100	9410	17800	7190	9520	2720	3030	2840
29	2400	13800	7170	4430	8650	9710	16200	8930	7900	2370	2920	2170
30	1290	14300	6960	4280	---	8580	15200	22100	7270	3590	2810	1460
31	3310	---	7040	4370	---	7820	---	38300	---	3870	2890	---
TOTAL	89569	325840	360420	154260	207390	251920	674720	409140	428040	167660	86550	75280
MEAN	2889	10860	11630	4976	7151	8126	22490	13200	14270	5408	2792	2509
MAX	4160	29400	38200	6380	13700	16500	59600	38300	48500	9580	3880	3250
MIN	769	3060	6470	4250	4030	4640	8240	7160	1020	2370	1940	1160

CAL YR 1983 TOTAL 2938165 MEAN 8050 MAX 39900 MIN 622
WTR YR 1984 TOTAL 3230789 MEAN 8827 MAX 59600 MIN 769

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LATE QUATERNARY STRATIGRAPHY OF THE LOWER ANDROSCOGGIN VALLEY, SOUTHERN MAINE

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INTRODUCTION

The late Quaternary history of the lower Androscoggin River valley is documented by complex stratigraphic relations between sediments of glacial, glaciofluvial, and glaciomarine origin. The entire coastal zone of southern Maine and nearby New Hampshire illustrates an environment that is unique to the New England region in terms of its Pleistocene stratigraphy and style of deglaciation.

The removal of the last, or late Wisconsinan, ice sheet from the coastal zone was accomplished by progressive marginal retreat of an active marine-based ice sheet. This is in contrast to other areas of New England where the retreating ice sheet was grounded on land and retreated by the progressive withdrawal of a stagnant ice margin (Koteff and Pessl, 1981). The <u>active</u> ice, retreating in contact with a shallow inland sea, deposited DeGeer, washboard, and stratified end moraines containing various diamicton facies and stratified drift below the limit of marine submergence (Bingham, 1981; Lepage, 1982; Smith, 1982, 1985; Thompson, 1982; Borns, 1986; Retelle and Konecki, 1986). Abundant outwash sediments were deposited in eskers, glaciomarine deltas, and submarine outwash fans beneath, and in front of the warm-based ice sheet. The coarse outwash facies grades distally (and vertically in cross-section) to fine-grained silty clay of the Presumpscot Formation (Bloom, 1960) that was deposited in the isostatically depressed foreland in front of the ice sheet. The fine-grained fossiliferous sediment blankets much of the surface of the coastal zone.

The purpose of this field trip is to examine stratigraphic evidence from a wide spectrum of glacial, glaciofluvial and glaciomarine environments on a transect from the coast, near Brunswick, to the inland extent of marine submergence near Poland Springs. We hope to visit exposures and sections that exhibit relations between various sedimentary facies and also point out some detailed information that we have obtained on the sedimentology and

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stratigraphy of these deposits. Discussion in the field will hopefully center on some of the factors such as water depth and subglacial topography (cf. Smith, 1984; Meier, 1985) that control the style and rate of ice retreat and the development of the stratigraphic sequence.

For a more comprehensive understanding of the glacial and glaciomarine stratigraphy of the region, the reader is referred to publications by Bloom (1960, 1963), Stuiver and Borns (1975), Smith (1982, 1984, 1985) and Thompson (1978, 1982). Maps of the surficial geology of the region have been completed at a reconnaissance level on scales of 1:24,000 and 1:62,500 Thompson and Borns (1985) have recently compiled the surficial geologic map of the state. More topic related research on endmoraine geomorphology and stratigraphy has been completed by students at the University of Maine at Orono (e.g. Bingham, 1981; Lepage, 1982; Attig, 1975) and Ohio University (Jong, 1980).

ACKNOWLEDGEMENTS

The authors express their sincere gratitude to the many people who have contributed to the production of the field trip. First, we would like to thank the gravel pit owners and operators, especially Mr. John Bisson and Mr. Ron Webber for allowing us to visit and study the geology of their property. Professors John Creasy, Roy Farnsworth and Donald Newberg from Bates College pointed out most of the stops we will visit on the trip and lent critical information and assistance throughout the study. Professor Joseph Hartshorn of the University of Massachusetts and Tom Weddle of the Maine Geological Survey visited the field sites and gave valuable critical assistance. Julie Retelle assisted in preparation of the field trip guide. Dr. Robert Stuckenrath of the Smithsonian institute kindly provided for a radiocarbon date for the Topsham site.

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ITINERARY
(See Fig. 1 for stop locations.)

<u>Mileage</u>

0.0 Topsham Fair Mall - assembly point. Brunswick 7.5 minute quadrangle
Leave Mall parking lot; turn right onto Route 196 (SE).

0.5 Turn left onto Second Street
Take an immediate left onto Route 201 (North) passing over Route 95 (1.4 miles)

1.8 Turn left onto Meadow Road. Pass over bedrock hill. Slow down at crest of hill.

2.5 Turn left into Bisson Pit.

2.9 End of pit access road.

<u>STOP 1</u>: <u>Bisson Pit</u>
This is an actively worked pit. Please pull off the access road to allow the trucks room to pass. This pit and the exposure at STOP 2

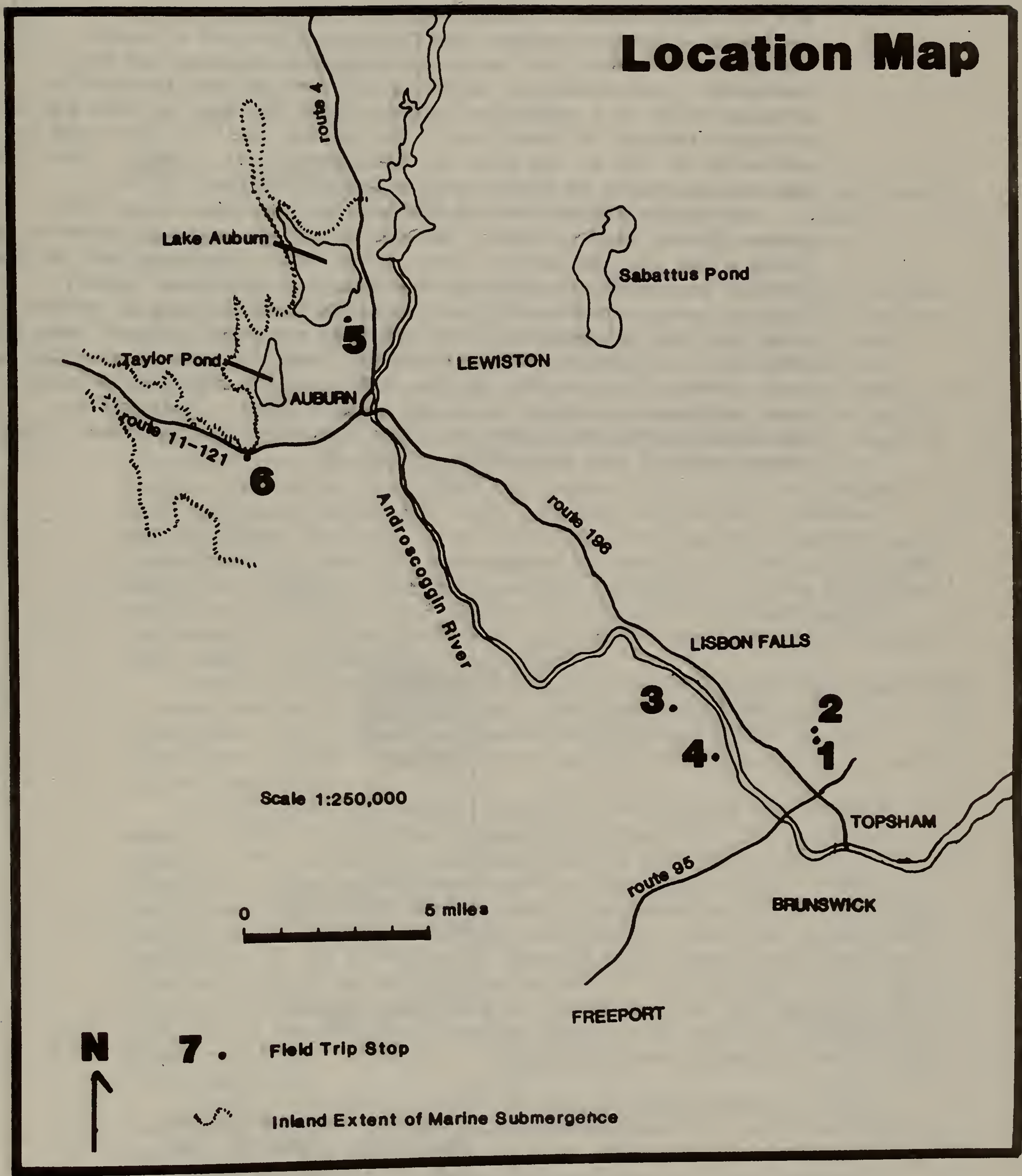

Location Map
route 4
Lake Auburn
Sabattus Pond
5
Taylor Pond
LEWISTON
AUBURN
route 11-121
6
Androscoggin River
route 196
LISBON FALLS
3.
2
4.
1
Scale 1:250,000
TOPSHAM
route 95
BRUNSWICK
0
5 miles
FREEPORT
N
7 .
Field Trip Stop
Inland Extent of Marine Submergence

Figure 1

are located along a bedrock ridge that trends in a northeast to southwest direction between the Cathance River and a linear topographic lowland that marks a structural lineament to the northwest. Approximately 5m of section are exposed through the proximal edge of a submarine outwash fan. The apex of this fan is at approximately 180 feet (asl); the marine limit for the area is estimated at 280 to 290 feet asl (Thompson et al., 1983). Thus, the fan was deposited in water depth of about 100 feet (30 m).

Sediments at the base of the section are cross-bedded to planer-bedded sandy gravel. Boulders as large as 1m in diameter are found along the ice contact head of the fan. The upper part of the section is medium to coarse sandy outwash. Individual gently dipping sandy outwash beds can be traced distally tens of meters from the head of outwash. Both gravelly and sandy outwash beds are faulted and folded in the ice proximal area. A compact gravelly mud unit (debris flow deposit or flow till) approximately 0.8m thick lies conformably within the sandy outwash unit. Laminated to massive Presumpscot Formation silty clay laps onto the lower lying distal side of the outwash fan. (Fig. 2)

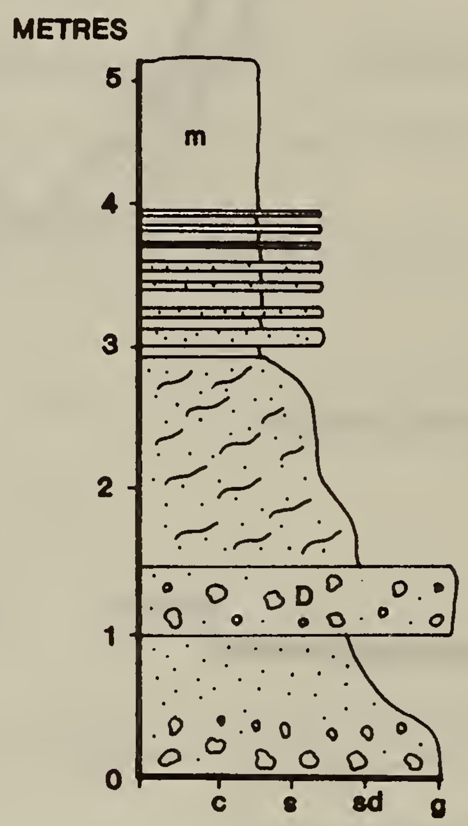


Figure 2. Composite stratigraphy of Bisson Pit - Stop #1 (see fig. 4b for column key)

Turn vehicles around and return to Meadow Road.

3.4 Turn left onto Meadow Road.

3.8 Turn left into Webber Pit.

STOP 2: Webber Pit

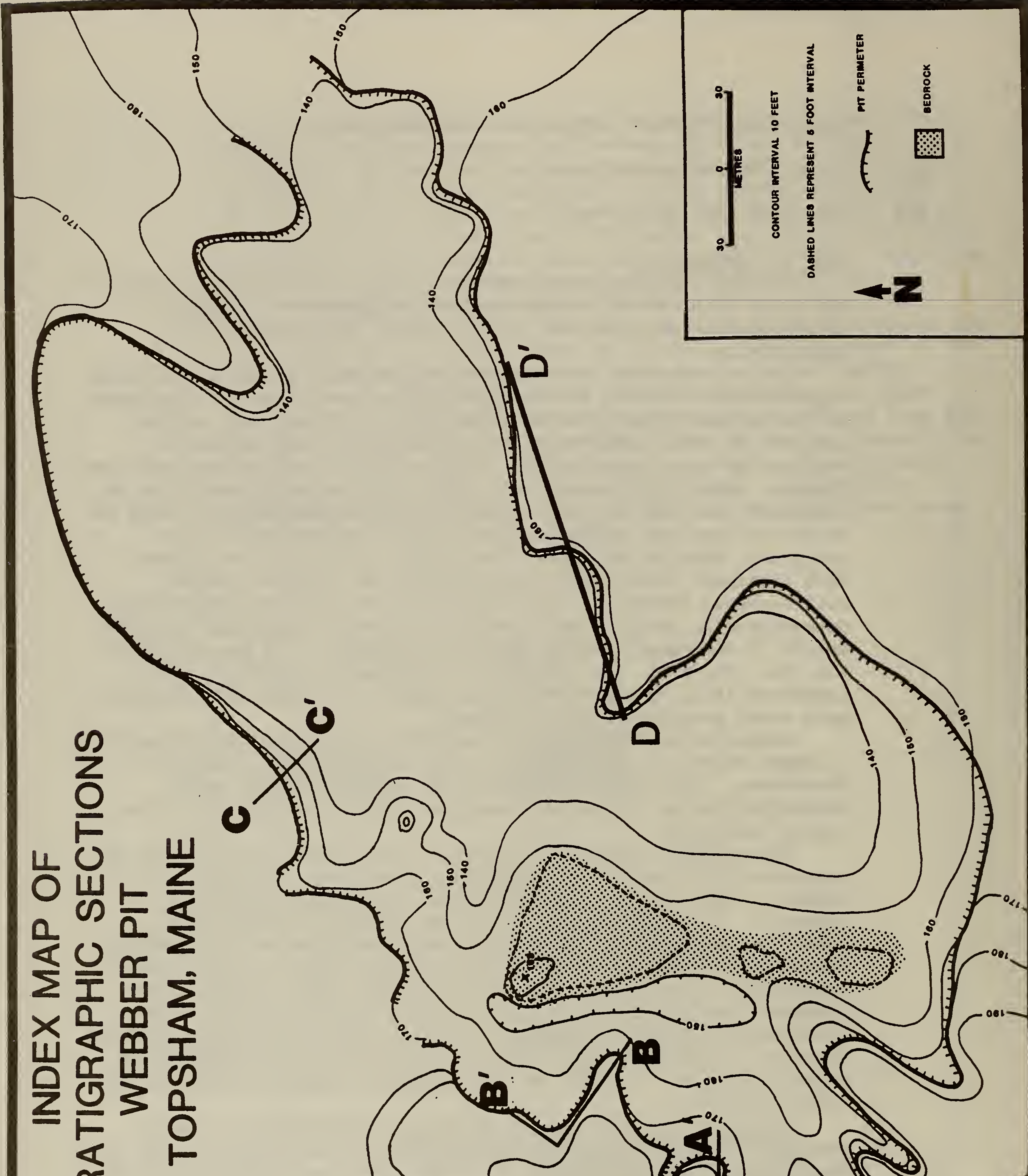
At this stop, depending upon available exposures at the time of the field trip, we will examine the stratigraphy and sediments interpreted as proximal and distal glacial and glaciomarine environments.

In the northwest and north-central sections of the pit several exposures are cut into a stratified end moraine (Fig. 3). The crest of the moraine ridge is at an elevation of 198 feet asl. approximately 100 feet (30m) below the marine limit. The moraine is composed of interbedded bodies of diamicton and stratified sands and gravelly sands. At least three layers of diamicton, consisting of lodgement till and flow till (resedimented lodgement till) range in thickness from approximately 50cm to 2m. Lodgement tills are texturally similar throughout the deposit (Fig. 4a), are compact with prominent lenticular sandy partings and contain lineated bullet-shaped boulders. The flow till is variable in texture and compactness in several exposures in the moraine depending upon the amount of reworking that the flow has undergone. In the middle of the flow till unit, the diamicton is similar in texture to the lodgement till; in exposures at the edge of the flow, reworking and more sandy interlamination is common.

Submarine outwash sands and gravels, most likely deposited from a tunnel source at the ice front, are interstratified with the diamictons. The outwash sediments are deformed by low angle thrustfaulting and recumbant isoclinal folding, especially at the top of the sands.

The proximal and distal flanks of the moraine are overlain by massive silty clay of the Presumpscot Formation. The silts are locally normal-faulted, presumably due to post-depositional slumping. Along the moraine crest and flanking outwash sands, poorly-sorted gravelly marine sands were deposited as the landform passed through wave base during post-glacial isostatic uplift and regression of the inland sea. Massive silty sands overlying the proximal zone of an outwash fan contain a rich in situ intertidal fauna consisting of a pavement of disarticulated *Mya arenaria* shells overlain by paired *Mytilus edulis* valves with attached *Balanus* in growth position. A sample of the *Mytilus* and *Balanus* dated 12,820 120; (SI-7017)

Along the southern margin of the pit approximately 6.5m of section is exposed through subaqueous outwash fan deposits (Fig. 3). Nearly all of the overlying 2 to 2.5m of the massive fossiliferous Presumpscot Formation has been stripped away by pit operations. Three outwash facies are recognized, primarily by the ratio of the thickness of sand to silt layers: (1) Proximal Subaqueous Outwash Fan Facies, (2) Distal Subaqueous Fan Facies and (3) Transitional Facies.


INDEX MAP OF
STRATIGRAPHIC SECTIONS
WEBBER PIT
TOPSHAM, MAINE
30
0
30
METRES
CONTOUR INTERVAL 10 FEET
DASHED LINES REPRESENT 5 FOOT INTERVAL
PIT PERIMETER
BEDROCK
N
A
A'
B
B'
C
C'
D
D'
140
150
160
170
180
190
200

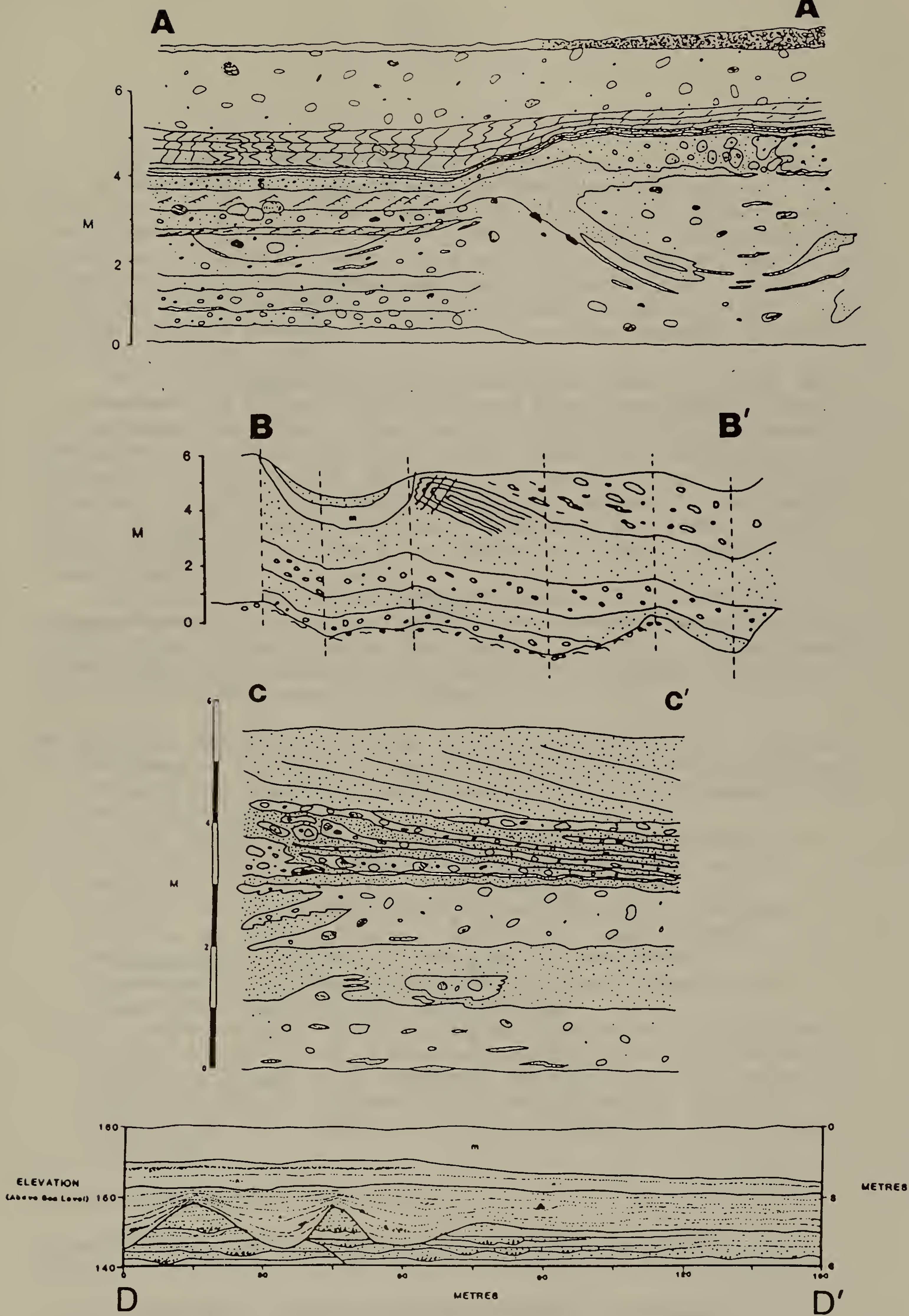


Figure 3. Index map and stratigraphic sections, Webber Pit, Topsham Maine.

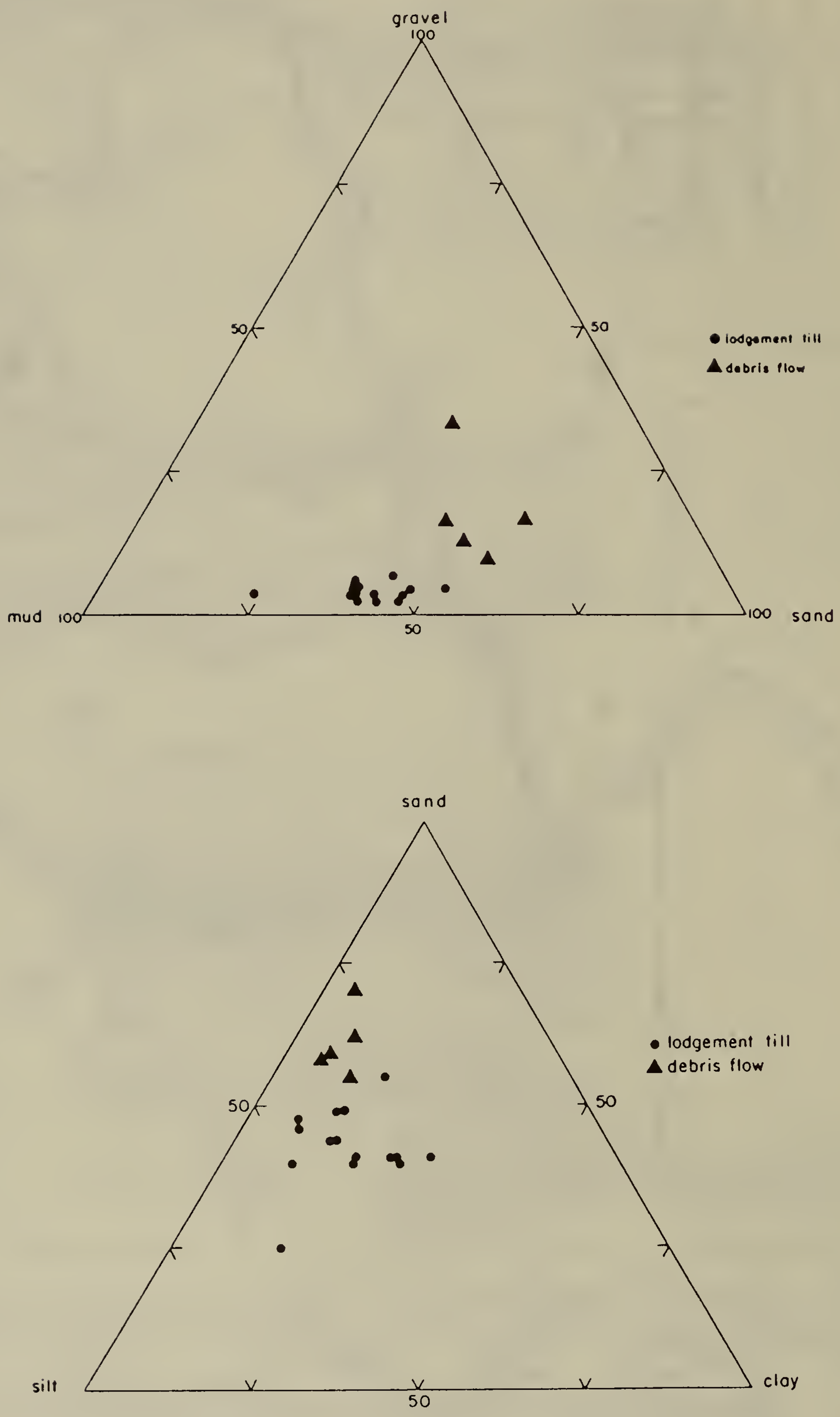


Figure 4a. Triangular plot of textural variations in diamicton facies in Webber Pit moraine (Stop 2). Gravel:sand:mud plot of 200 g samples. Sand:silt:clay plot of matrix.

The lowermost Proximal Subaqueous Fan Facies (Unit I in Fig. 4b) is characterized by sand layers thicker than silt layers. Truncated outwash fan lobes ranging 10m laterally and 3m vertically are evident. This facies is also characterized by: interbedded and graded beds of gravels, sands and muds; channel scour and lag deposits; ripple cross-laminations and climbing ripples. These features, and others which distinquish Facies II and III, are similar to those discussed in detail by Rust and Romanelli (1975) and Domack (1983, 1984).

The Distal Subaqueous Outwash Fan Facies (Unit II in Fig. 4b) is characterized by sand layer thickness being approximately equal to silt layer thickness. These layers are rhythmically laminated and graded silty sands and silty clays. Individual layers are traceable for at least 100 metres and the amount of mud increases laterally. These layers thicken within the channels cut through Unit I and soft sediment deformation structures are common. Ice rafted dropstones and dropped (or slumped in) diamict masses are found frequently.

The Transitional Facies (Unit III in Fig. 4b), which grades into the massive Presumpscot Formation is characterized by the silt layers being thicker than the sand layers. The thinly bedded graded silty clays intertongue with silty sand; soft sediment deformation and ice rafted dropstones are present.

We summarize the glacial geology at this site as follows: (Fig. 5)

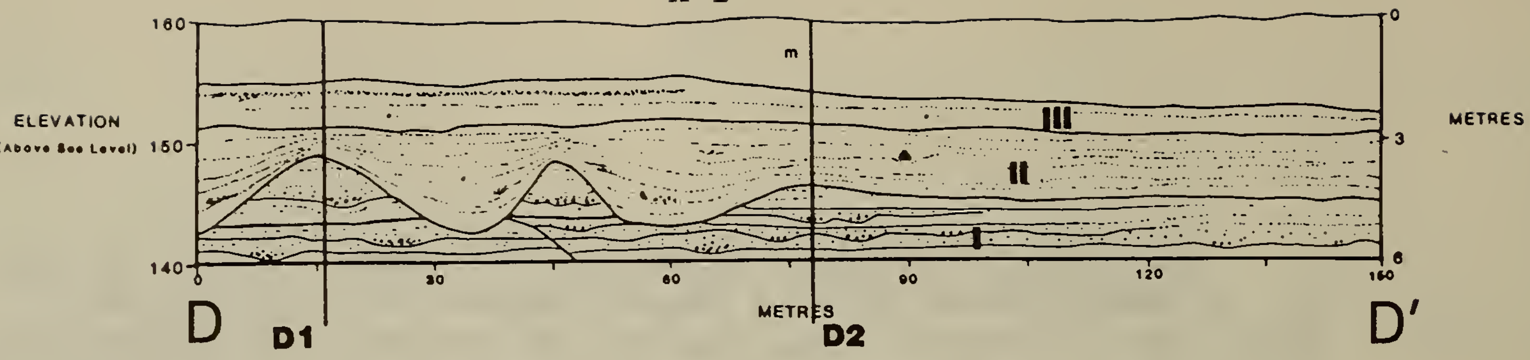
1. Progressive ice marginal retreat from south to north this area is marked by successive submarine deposited in approximately 100 ft. (30m) water depth.
2. The ice margin became grounded or pinned when it retreated to the high point of the bedrock ridge underlying the western section of the Webber Pit. The stratified moraine was constructed by the oscillation of the ice margin. Lodgement till was deposited during stillstands or slight forward advances, perhaps during winter when calving ceased. Interstratified outwash beds were laid down during the melt season and later overridden and deformed.
3. Ice retreated more rapidly to the east where the ice front was not pinned against the bedrock topographic high. Accelerated calving in deeper water influenced the rapid retreat. Submarine outwash sequences, exposed in the southern margin of the pit fine upward and demonstrate the increasing distal nature of the sediments.

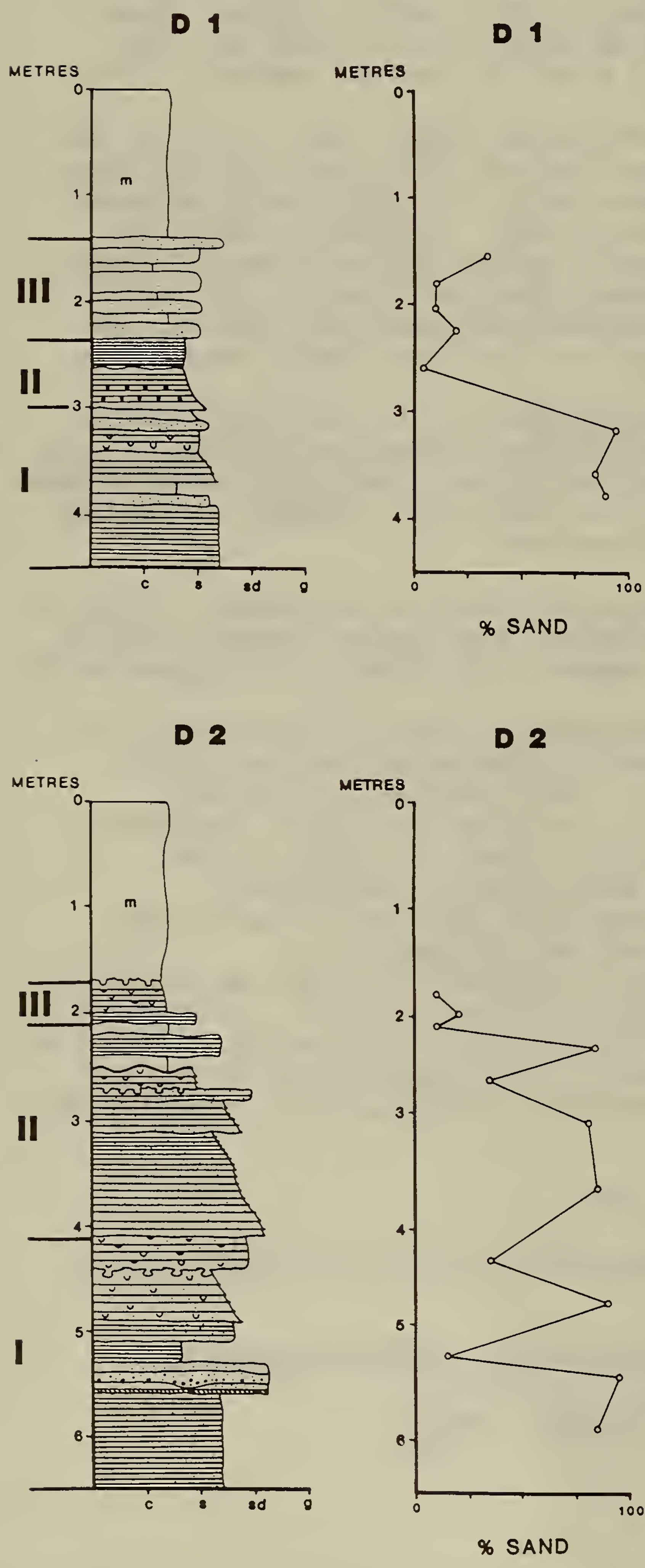
4.0 Return to Meadow Road, turn left.

4.9 After crossing swampy lowland, turn sharp left onto Cross Meadow Road.

5.9 Cross brook, enter Lisbon Falls South 7.5 minute quadrangle.

6.9 Intersection, Route 196; turn right (west).





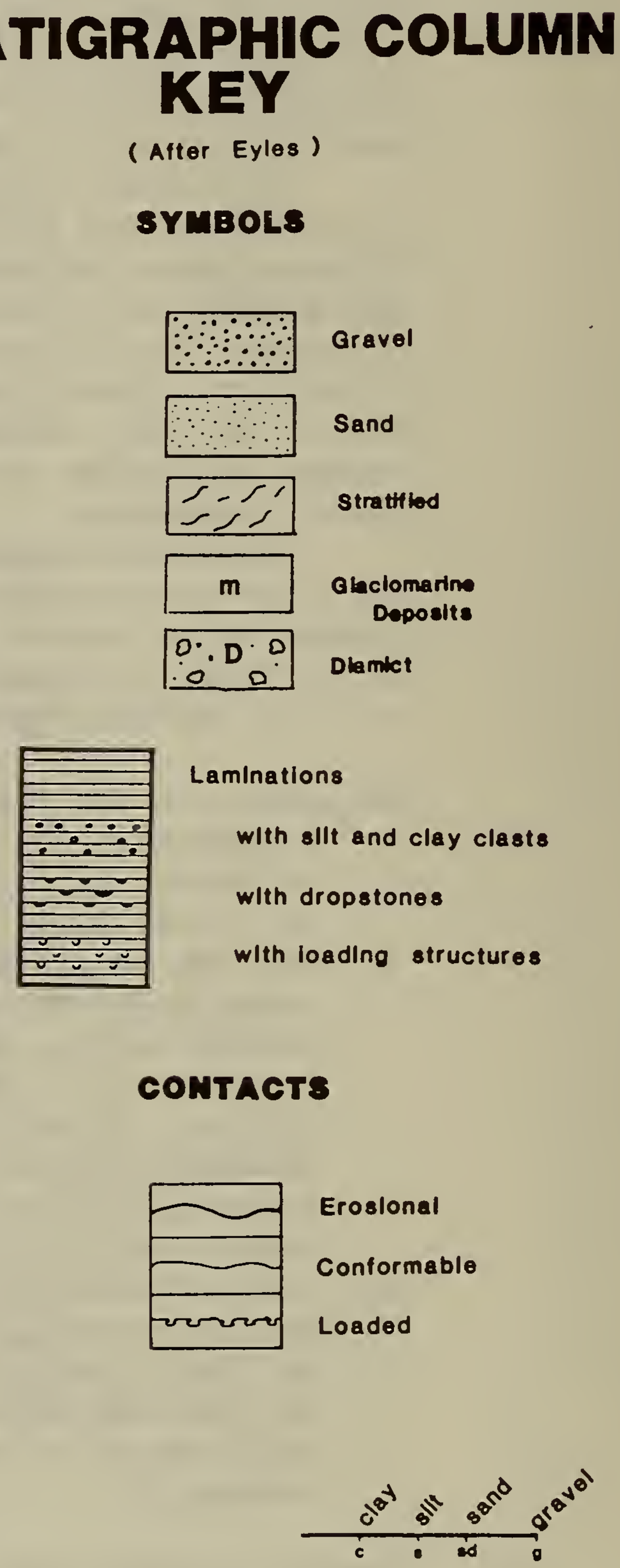


Figure 4b. Stratigraphic section and vertical profiles, Webber Pit (Stop 2).

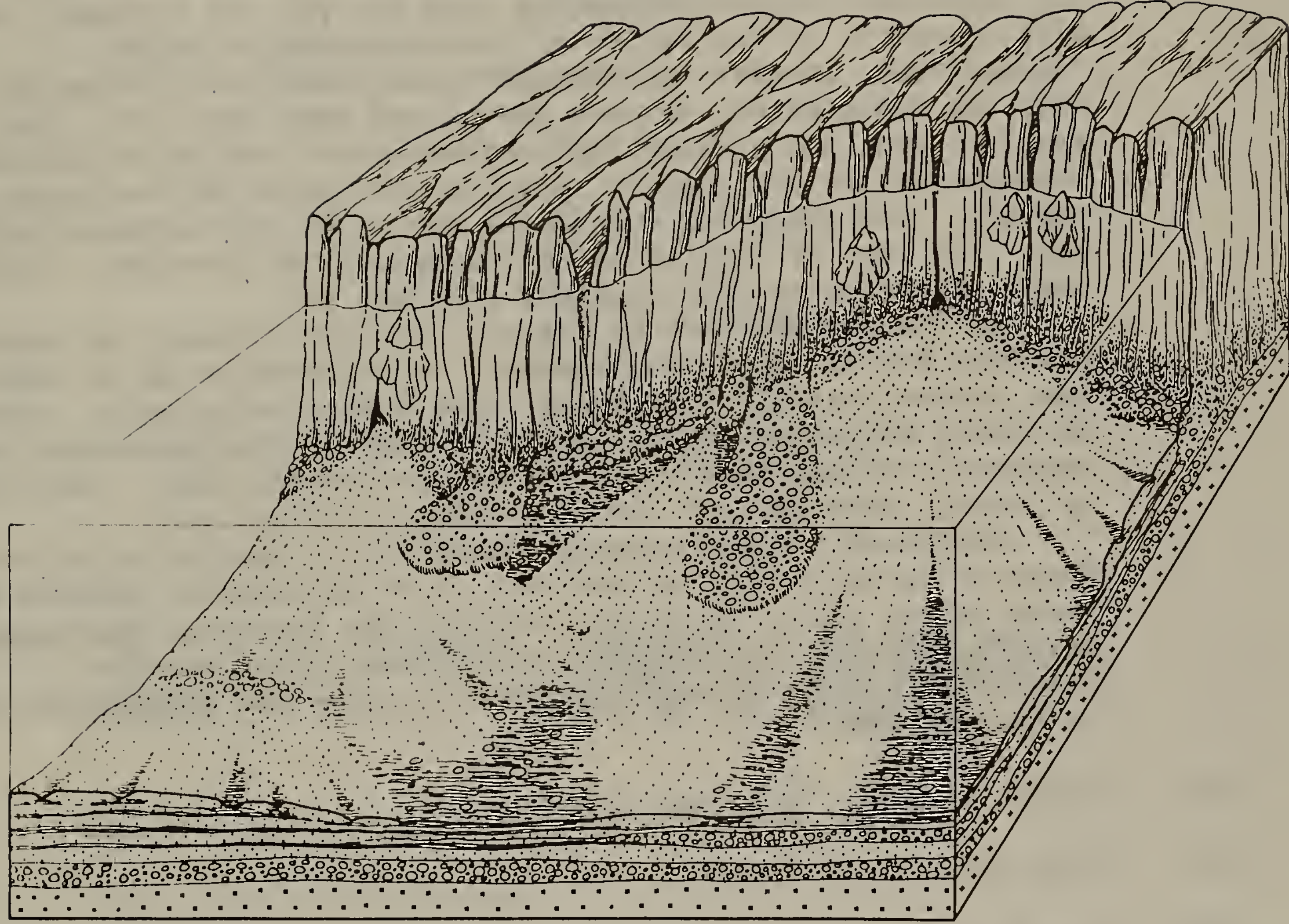


Figure 5. Diagrammatic model of glacio-marine deposition based on exposures at Webber Pit, Topsham, Maine.

10.5 Androscoggin County Line, Village at Lisbon Falls.

11.0 Junction Routes 9 and 125. Turn left.

11.3 Cross Androscoggin River, turn left at end of bridge.

11.4 Follow Route 125 passing large open gravel pit then uphill.

12.2 Turn left into Tupper Construction Pit.

STOP 3: Tupper Gravel Pit, Durham

We will park on the pit access road and proceed to an overlook for discussion before descending into the pit for a closer look. At this exposure, in the center of the Androscoggin Valley, fine-grained, massive and fossiliferous Presumpscot Formation overlaps ice-proximal outwash gravel and sand (Fig. 6). The glaciofluvial and glaciomarine sediments are incised by fluvial deposits of the post-glacial Androscoggin River on the north wall of the pit. The upper surface of the pit is at approximately 200 feet asl. The base of the pit is approximately 170 feet asl. The marine limit elevation is approximately 310 feet asl.

As the eastern base of the pit, a linear deposit of cobbley to bouldery gravel (40-50 cm diameter) is interpreted as an esker ridge that grades distally to finer grained submarine outwash. Overlying the esker ridge are several gravelly sand to sand sequences that represent the aggradation of successive outwash fans. The outwash is locally deformed and contains bodies of flow till.

Laminated to massive-bedded marine silt laps onto the outwash sequence in the center of the pit. Rich macrofaunal remains are found in the silt. Pelecypods include *Mya truncata*, *Mya arenaria*, *Hiatella arctica*, and *Mytilus edulis* among other species. Additionally several large gastropod species and *Balanus* are found in the sediments.

12.7 Return to Route 125; turn left.

14.1 Turn left on Soper Road.

15.9 Turn left into pit (Cianbro Corp. sign).

16.2 Park in the bottom of the Pit.

STOP 4: Chick Pit

This pit is located approximately 1/2 mile southeast of a large northeast trending moraine and within the pit a bedrock ridge trends northeast (Chick 1986). The upper surface of the pit is at approximately 200 feet asl and the local marine limit is approximately 310 feet asl. Thus the materials were deposited in water depth of approximately 100 to 120 feet (30-40 m). Good exposures within two sections of this pit show a series of overlapping outwash fan deposits topped by 2 to 3m of Presumpscot Formation. Several outwash fan heads are visibly recognized by the location of boulder concentrations, deformation within the outwash deposits, and surface topographic expression of the fans (lobate form).

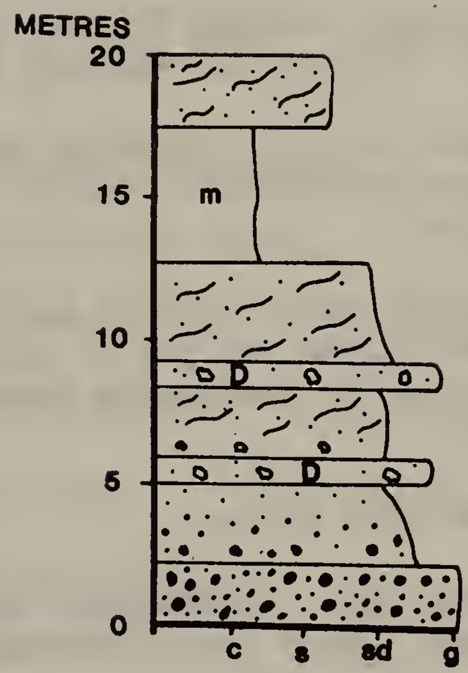


Figure 6. Composite stratigraphy of Tupper Pit - Stop #3 (see Fig. 4b for column key)

Turn vehicles around and return to Soper Road.

17.3 Turn right onto Soper Road.

19.1 Intersection with Route 125. Proceed straight ahead on Soper Road.

19.5 Stop sign. Intersection of Meadow Road. Proceed straight ahead.

21.5 Intersection Route 9. Stop sign. Proceed straight ahead on Route 9 (west).

22.8 Enter North Pownall 7.5 minute quadrangle.

22.9 Intersection with Route 136. Take right turn (north).

23.7 View across Androscoggin River to right. Broad river terraces on left.

24.6 Enter Lewiston 7.5 minute quadrangle.

26.4 Rise upon terrace.

26.5 Auburn townline.

30.0 Cross under Maine Turnpike.

31.9 Y-Intersection. Follow Route 136 left on Mill Street.

32.0 Broad Street straight through lights. At 2nd stop light follow Route 136 right. Stay on 136 through 3rd and 4th lights.

32.8 Junction Route 202/100/11. Continue straight across intersection (Court Street) to LaVerdiere's Drug Store. Take left in front of store at stop sign.

33.0 Turn right at light; keep to right and proceed ahead on Route 4. There are several places for you to purchase lunch, either sitting in a fast food place or purchasing groceries to eat at the lunch stop.

33.5 Beginning of Fast Food strip.

34.4 Intersection Route 202. Auburn Mall on left, continue straight ahead. Enter Lake Auburn East quadrangle.

36.1 Lunch stop and reassembly point. After lunch we will turn around and proceed south on Route 4. Take right turn onto Turner Street.

36.3 Lido gas station.

37.4 Intersection with Grace Lawn Road. Take right turn.

37.7 Right turn into Auburn Landfill.

STOP 5: Auburn Landfill - Gravel Pit

There are several exposures at this site that expose the topset, foreset, and bottomset beds of a glaciomarine delta. The north slope of the landform is kettled and exhibits good evidence of collapse. Pegmatite bedrock crops out along this slope and above the topset plain. The highest elevation on the topset plain is 360 feet elevation. The topset-foreset contact, measured at 336 feet asl. (Thompson et al., 1983) is illustrated by a spectacular cobble and boulder horizon in the north central area of the pit. Sandy foreset beds prograde southward over massive sandy silt bottomset beds from which several whole valves of Hiatella arctica were recovered.

This landform, which is graded to relative sea level of 336 feet is in the form of a classic Gibert type delta, is distinct among the fluvial marine features that we have examined already today. Other sandy fluvial beds were depostied on fans that graded to the sea floor at a depth of approximately 100 feet (30m). Perhaps this deposit originated as a submarine fan and later aggraded to sea level due to a constant sediment source with the ice sheet pinned on the bedrock topographic high.

Turn right on Grace Lawn Road leaving the landfill.

38.2 Enter Minot 7.5 minute quadrangle. Mt. Auburn Cemetary on left.

38.5 Turn left on Park Avenue.

38.6 Cross Summer Street at stop sign. Proceed straight ahead on Park Avenue.

39.4 Enter Lewiston 7.5 minute quadrangle.

39.7 Stop sign. Proceed straight ahead.

40.4 Intersection with Court Street. Turn right.

41.4 Intersection with Minot Avenue. Turn right at light.

41.7 Intersection. Stop light. Go straight ahead.

42.8 <u>STOP 6</u>: <u>Minot Road Borrow Pit</u>

Please pull off the road into the pit on the left (southside of Minot Road). Watch for oncoming traffic around the corner as you cross the left lane.

This pit is located on the southern end of a till/bedrock hill (Mt. Apatite) at an elevation of 300 feet asl. While this site is probably below the marine limit (approx. 340 feet, Thompson et al., 1983) parts of the hill to the south of the borrow pit are above the estimated marine limit elevation.

Several small exposures in the pit are cut into a sandy silty diamict that is distinctly different from other till or debris flow deposits that we have already seen today. At this site there is approximately 2.5 to 3.0m of sandy till that overlies striated pegmatite bedrock (N 25 W). The till has a distinctive sub-vertical to vertical fabric seen in fissility, in sandy interlaminations and in till clasts which is exposed in several areas of the pit. The "till" is texturally and structurally similar to the "upper till" of southern New England (cf. Koteff and Pessl, 1985) and the sandy drift at New Sharon (T. Weddle, pers. comm., 1985). The till is also similar to that in several exposures in the uplands about one mile northwest of the pit. The section is capped by 50 to 70cm of eolian silty sand that overlies a ventifacted pavement on the till ridge surface.

The genesis of both the sediment and the landform at this site is not well understood and we invite your discussion.

This is the formal end of the trip. If time allows, we offer you an alternative return route with one quick stop in an esker to view stratigraphy and structures in an ice contact setting. We estimate approximately 1 hour for the extended trip. If you wish to return to the Lewiston-Auburn area, turn right out of the pit and follow Minot Avenue east towards Auburn. To get to downtown Lewiston-Auburn and the Bates College area, follow Minot Avenue (Route 11-121) to the intersection with Central Avenue, bear left at intersection. Follow Central Avenue (Route 4/100) north to downtown Auburn. To return to Bates College, turn right on Court Street (lights) and cross Androscoggin River. Go uphill approx. 1/2 mile, bear right onto Sabattus Street and take the second left onto College Street. Follow until you reach the campus. For those continuing on the trip a handout sheet will be provided with additional instructions.

MID-PALEOZOIC CALC-ALKALINE IGNEOUS ROCKS OF THE NASHOBA BLOCK AND MERRIMACK TROUGH

Rudolph Hon[1], J. Christopher Hepburn[1], W. A. Bothner[2], William J. Olszewski[2], Henri E. Gaudette[2], William H. Dennen[3], and Christer Loftenius[1]

(1) Department of Geology and Geophysics, Boston College, Chestnut Hill, MA 02167
(2) Department of Earth Sciences, University of New Hampshire, Durham, NH 03824
(3) Nahant, Massachusetts

INTRODUCTION

In northeastern Massachusetts and adjacent southeastern New Hampshire there are a dozen or so undeformed bodies of norite, diorite, quartz diorite, and granodiorite with associated volcanics of similar composition. These occur within three different tectonic zones. The igneous rocks all yield ages between 470 and 400 Ma. From southeast to northwest the tectonic zones are the Newbury Inlier, Nashoba Block, and Merrimack Trough of Lyons and others (1982). This triplet lies between the Avalonian Terrane (Boston Block) to the SE and the Acadian Kearsarge-Central Maine Synclinorium to the NW. Our intention is to focus on the comparison, origin and tectonic significance of the intermediate igneous rocks within this triple tectonic set.

NEWBURY INLIER

This small tectonic wedge is most significant in that it contains fossiliferous strata and significant occurrences of bimodal andesite-rhyolite volcanic suite. Hon and Thirlwall (1985) presented geochemical evidence suggesting that the andesite formed in a subduction zone environment. If their conclusion is correct then the Newbury Inlier would indicate the existence of a subduction zone from at least the Upper Silurian to earliest Devonian time. This subduction would have ceased at the beginning of the Acadian deformation, and thus may be critical to our understanding of the nature of that orogeny.

NASHOBA BLOCK

Within the Nashoba block we recognize three or four bodies which have a geochemical character similar to the Newbury volcanics (see Fig. 1). These are the Preston gabbro, Straw Hollow Diorite, Assabet Quartz Diorite, and Sharpner's Pond Diorite (Fig. 2). For more detail see the roadlog section under the individual stops.

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(2) Department of Earth Sciences, University of New Hampshire, Durham, NH 03824;
(3) Nahant, MA 01908

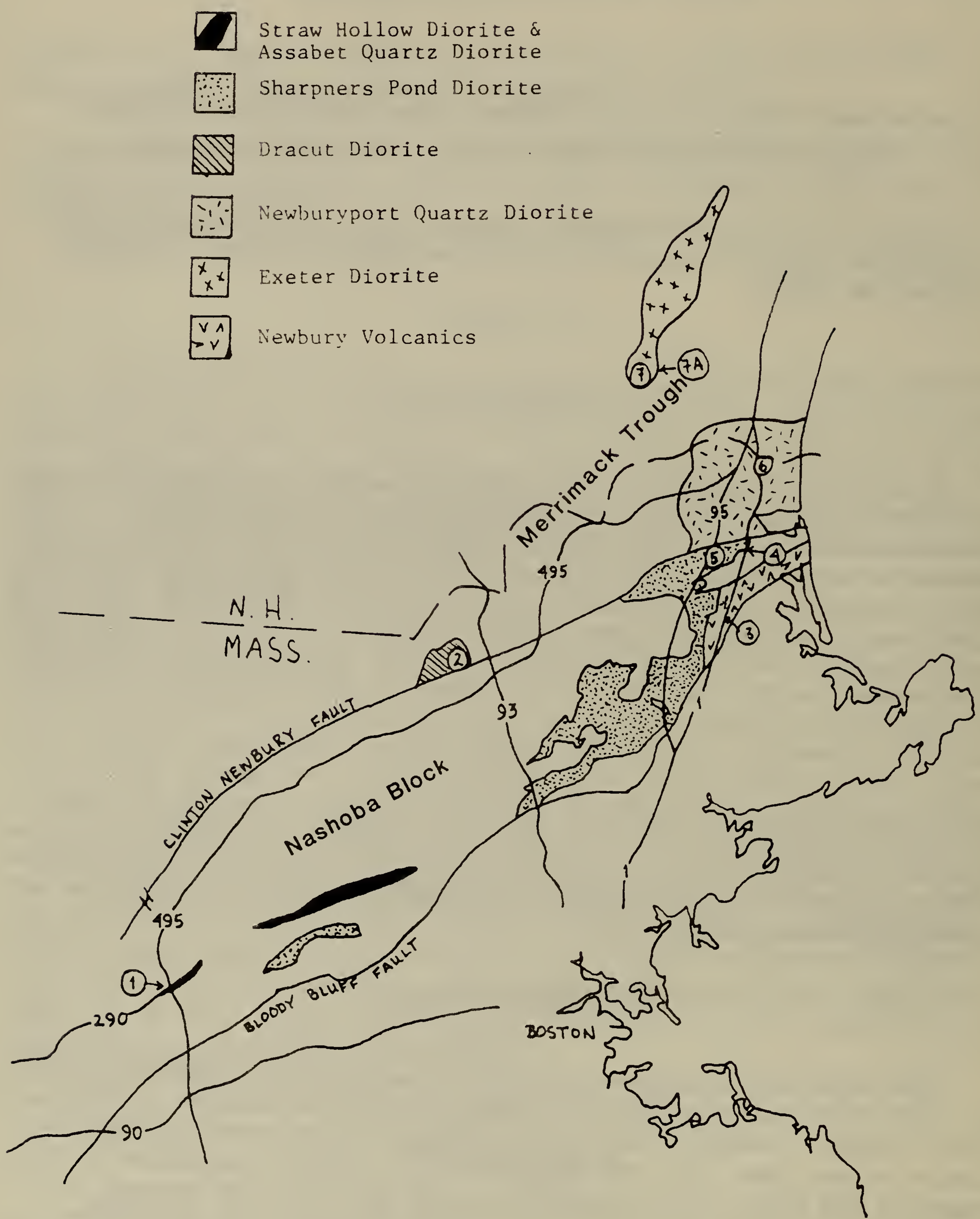
INTRODUCTION

In northeastern Massachusetts and adjacent southeastern New Hampshire a dozen or so, undeformed or little deformed calc-alkaline plutonic bodies and volcanics of similar age are found within a region which spans over three different tectonic zones. Separated from each other by regional faults they include the Newbury Inlier to SE, the Nashoba Block in the middle, and the Merrimack Trough of Lyons and others (1982) to NW. This "triplet" itself then lies between the Avalonian Terrane (Boston Block) to SE and the Acadian Central Maine - Kearsarge Synclinorium to NW. The fault system which bounds the Merrimack Trough on the southeast is the Clinton-Newbury Fault Zone, and the fault, which separates the Nashoba Block from the Newbury Inlier, is the Parker River Fault Zone. The composition of these calc-alkaline plutonic bodies varies generally from norites, diorites, quartz diorites, to granodiorites, and the volcanics are high alumina basalts, basaltic andesites and andesites. Associated with each of the individual suites are also prominent granitic/rhyolitic rocks which appear to be closely related in time but unrelated by a magmatic lineage. Our trip will focus on a comparison of the origin and tectonic significance of the intermediate igneous rocks, all of which yield distinct pre-Acadian ages between 400 and 470 Ma.

NEWBURY INLIER

This small tectonic wedge (see also Bedrock Geologic Map of Massachusetts, Zen, ed., 1983) caught along the Bloody Bluff Fault Zone is most noted for its Silurian to Lower Devonian fossiliferous strata and for significant occurrences of bimodal andesite-rhyolite volcanic suite. Hon and Thirlwall (1985) presented geochemical evidence suggesting that these andesites formed in a subduction zone environment. If their conclusions are correct, then the Newbury Inlier would indicate the presence of a subduction zone, which lasted for a minimum from Silurian to Lower Devonian and ceased just prior to the onset of the Acadian deformation.

FIGURE 1. Stop locations.



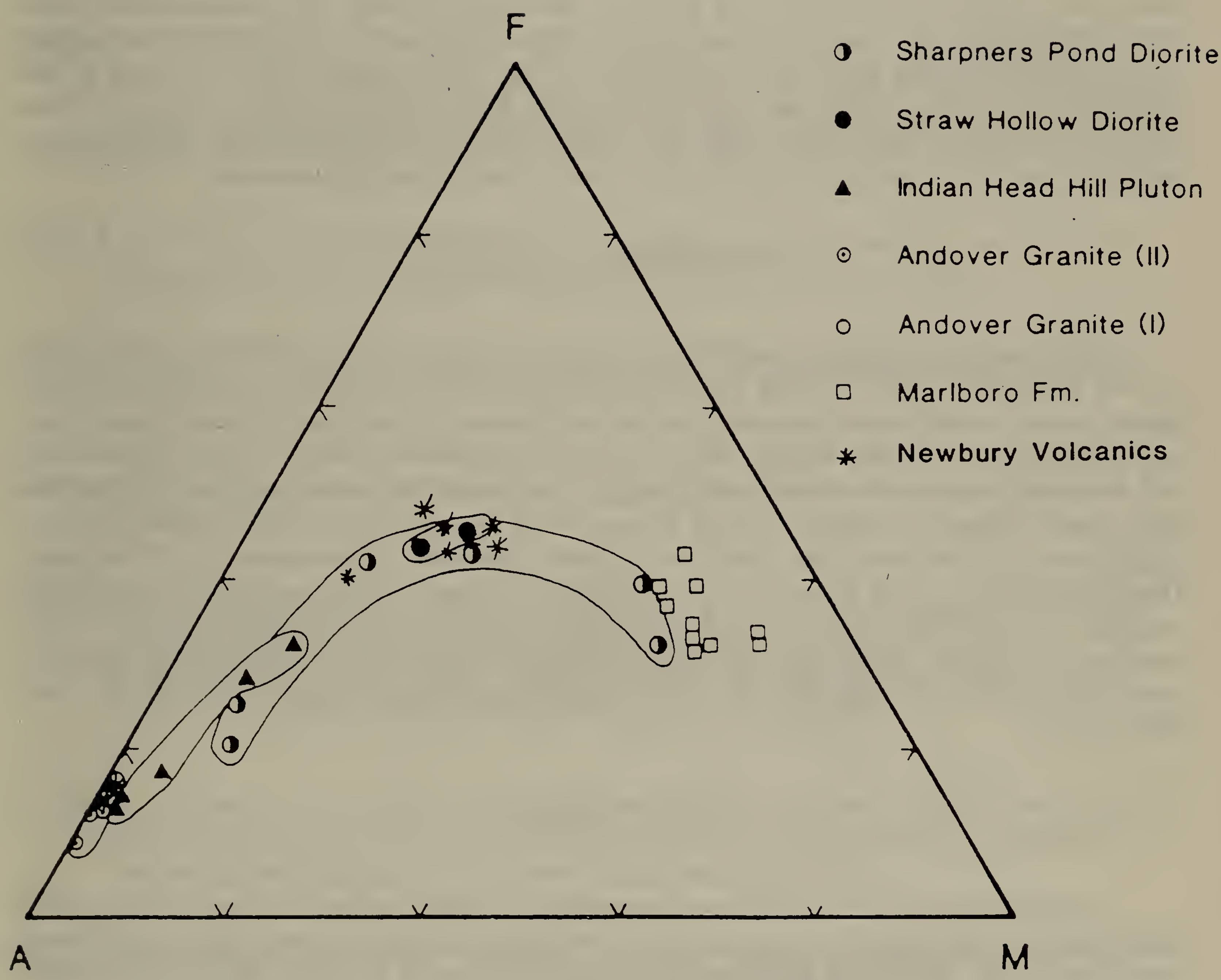


FIGURE 2 AFM Diagram Nashoba Block.

NASHOBA BLOCK

About half the block is underlain by little deformed or undeformed calc-alkaline plutonic rocks and by contemporaneous peraluminous and metaluminous granites yielding radiometric ages of 402 to 455 Ma. The rest of the Nashoba block consists of stratified rocks which are mostly metamorphosed volcanics and volcanogenic sediments of Lower Paleozoic to Upper Proterozoic age (pre-455 but post-730 Ma). The intermediate rocks found throughout the Nashoba Block form variable sized and elongated plutons aligned conformably with the regional trend (Fig.1.). The calc-alkaline trend of some of the analyzed intermediate plutonic rocks (Sharpners Pond Quartz Diorite, Straw Hollow Diorite, and Indian Head Hill) is shown on Fig. 2. which also has plotted peraluminous Andover Granite, metamorphosed Marlboro basalt, and Newbury andesites.

MERRIMACK TROUGH

Only recently has Merrimack Trough been recognized as a possible separate terrane (Lyons and others, 1982). More work, of course, needs to be done on these rocks but several recent contributions indicate, that this region may indeed be a separate terrane (see for example: Bothner and others, 1984; Gaudette and others, 1984). A rather interesting and to some extent even tantalizing suggestion made by these authors is, that the Merrimack Trough is actually a Late Precambrian Block which, on the whole, did not significantly suffer during the Acadian Orogeny. However, what makes the Merrimack Trough interesting, from the perspective of this excursion, is the presence of considerable volume of calc-alkaline rocks, which are typically associated with subduction zones near continental margins. If the ages of these intrusions are allowed to be used to time the presumed subduction zone, then the subduction lasted from around 470 Ma ago to about 400 Ma ago.

CONCLUSIONS

At present, not enough data are yet available to fully assess the significance of the mid-Paleozoic calc-alkaline magmatism in northeastern Massachusetts and southeastern New Hampshire. Several M.Sc. theses currently underway at Boston College may perhaps provide some of the key data in the near future.

In conclusion, we would like to suggest that the mid-Paleozoic calc-alkaline suite may be indicative of a long lasting (appr. 70 Ma) subduction zone which preceeded and might have ultimately led to a continental collision - the Acadian Orogeny. The position of the arc axis probably was not at the same place all the time but most likely migrated from place to place leaving behind plutonic bodies of discrete ages. It would be premature at this time to argue for the exact position or the polarity of this subduction zone, but the essence provided by the mid-Paleozoic calk-alkaline suite is, that such a subduction zone might

have existed and that it was most likely responsible for the high temperature regime causing the widespread crustal anatexis.

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ROADLOG FOR MID-PALEOZOIC CALC-ALKALINE IGNEOUS ROCKS OF THE NASHOBA BLOCK AND MERRIMACK TROUGH.

ITINERARY

Mileage: Part I.

0.0 Assembly point and Stop #1. Meet at junction of Interstates I-290 and I-495 just north of Marlborough, Mass. at 10 a.m. Meet on S.E. side of this large interchange. If coming north on I-495, take exit 25A "to 85, Marlborough", park on shoulder of ramp just before it merges with I-290 by silver colored electric box. If coming east on I-290, continue over I-495 and park just east of where the ramp from I-495N, joins I-290 extension "to 85, Marlborough". We will leave this exposure at 11 a.m.

STOP #1: STRAW HOLLOW DIORITE or ASSABET QUARTZ DIORITE
The large exposures at this cloverleaf are complex and show a wide variety of features. Our principal purpose in stopping here is to examine the Straw Hollow Diorite or Assabet Quartz Diorite along the north side of the interchange. The diorite here is characteristic of the smaller bodies of Silurian calc-alkaline plutonic rocks in the Nashoba Block. Foliated and nonfoliated, or more weakly foliated varieties of hornblende-quartz diorite are present here. They have been intruded by garnet-bearing pegmatites associated with the younger phase of the Andover Granite (approximately 415 m.y., Hill et al., 1984). Sillimanite schists of the Nashoba Fm. are exposed along the south side of the cut. Blastomylonites believed to be associated with the Assabet River fault zone are also exposed here and will be briefly examined as an example of the type of deformation common along some of the larger shear zones in the Nashoba Block. If time permits, a relatively late brittle shear zone cutting the Straw Hollow, with carbonate mineralization, will be visited to provide contrast with the earlier ductile deformation.

To proceed to Stop 2 we will need to turn around and head north on I-495. Continue straight ahead on "to 85" east.

1.0 At traffic light, Fitchburg Street, exit to the right.

1.1 By entrance to Assabet Valley Regional Vocational High School, turn around and retrace route to traffic light.

1.2 Turn left (west) on "to 495-290".

2.2 Take Exit 26B, I-495 north toward Lowell.

2.8 Merge with I-495 north.

3.0 Cross Assabet River.

3.8 Excellent exposures of Nashoba Fm. gneisses and migmatites for next several miles, continue north toward Lowell.

25.8 Rest area on right. Odometer check.

26.7 Leave I-495 at Exit #36, "Lowell Connector". Exit to right.

27.4 Turn right, follow Exit 36, Lowell Connector.

28.1 Continue straight on Lowell Connector.

28.8 Continue toward Lowell Center.

29.9 Road ends, turn left toward Lowell Center.

30.1 Continue straight past Courthouse on right.

30.2 Bear right, follow one way.

30.4 Bear left, continue on one way.

30.45 At light, continue straight, now on Central St.

30.6 At light, Y intersection, bear right on Prescott St.

30.7 At light, turn right on E. Merrimack St.

30.8 Federal Building on left.

31.0 Continue straight.

31.1 At light, turn left on Routes 38 and 110.

31.3 Cross Merrimack River and bear right on Route 110 east, you are entering the Merrimack Trough.

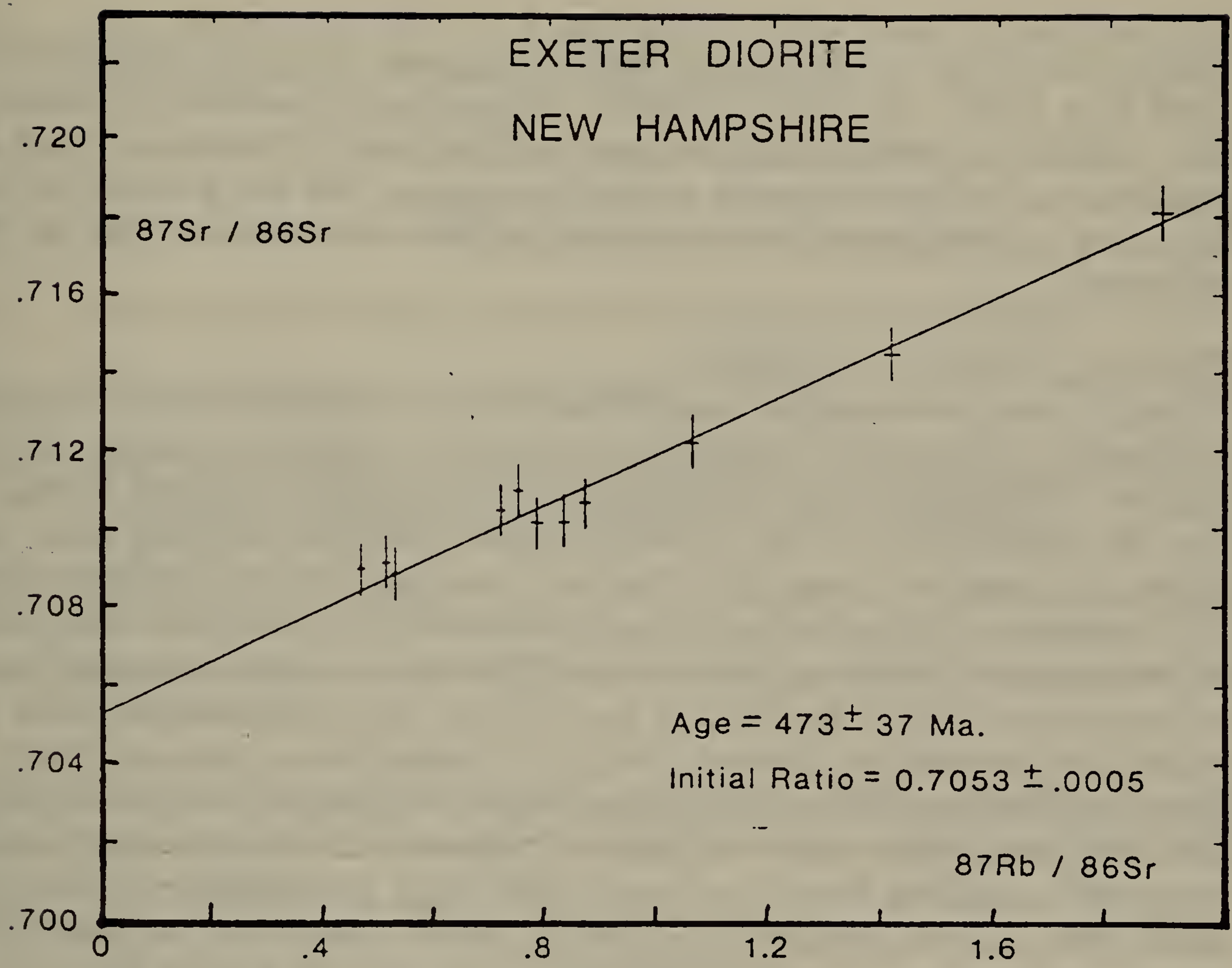
NOTE: If you should for any reason deviate from the Roadlog, find your way to Route 110 east on the northern side of Merrimack River and proceed 4.0 miles toward the turn-off to George Brox, Inc. quarry.

31.4 Continue on Route 110 east toward Lawrence.

32.2 Dracut town line.

34.9 Cross under power lines.

35.3 Turn left onto side road at sign for George Brox, Inc., proceed up hill. WATCH FOR TRUCKS!



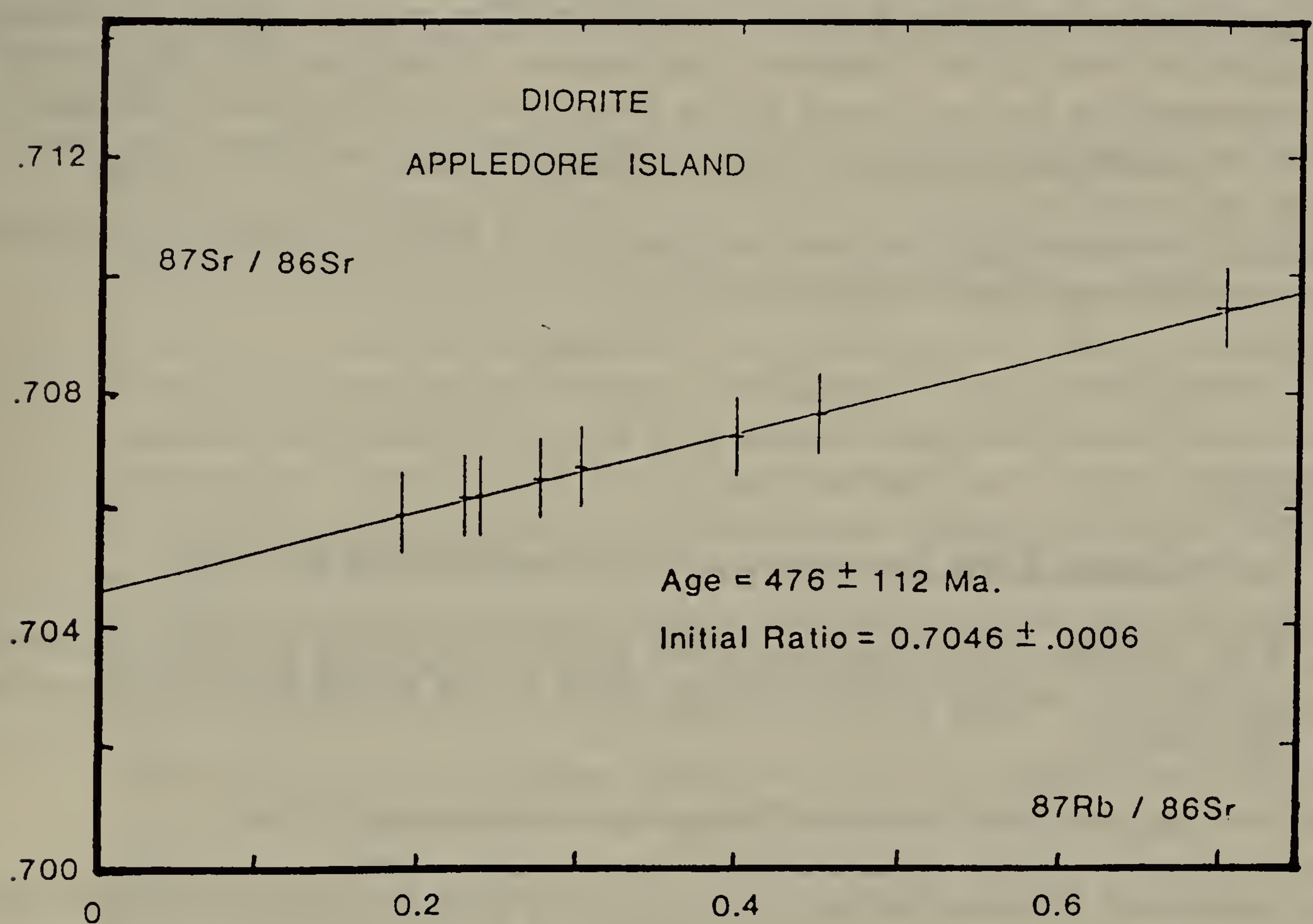


FIGURE 3 Rb/Sr Isochrons for Exeter and Appledore Island Plutons

Rb Decay Constant = 1.42×10^{-11} Years.

35.7 Park in lot on left at entrance to Brox, Inc quarry. WATCH FOR TRUCKS-THIS IS A VERY ACTIVE QUARRY. At this locality we will consolidate to as few vehicles as possible and proceed into the quarry. Prepare to stop at the gatehouse and obtain clearance to enter the quarry. We will proceed driving to the lower level and park in an area which does not interfere with the quarry operations.

STOP #2: DRACUT DIORITE.
Virtually every petrographic type known to occur within the Dracut stock can be found in this quarry. The main rock type is a massive, medium- to coarse-grained, unfoliated norite. Petrographic studies done by Dennen (1942) show that the mineralogy of the Dracut is highly variable ranging from noritic diorite to cumulate ultramafics. The latter consists of 55% forsteritic olivine, 20% enstatitic orthopyroxene, 5% each bytownite and clinopyroxene, and 15% of intercumulus biotite and hornblende. The norite contains subequal amounts of plagioclase (An71-An37) and mafics (cpx, opx and olivine) with minor biotite and hornblende. Dennen (1942,1943) describes a gradation from more felsic varieties near the contacts to more mafic norites in the center, noting that this zonation is likely due to magmatic fractionation. A curious occurrence of magmatic sulphides which can reach quite high proportions in some places (over 50%), led to a prospecting effort for nickel - hence Nickel Mine Hill, the location of this quarry. The Dracut crops out over an area of about 27 mi^2, intruding rocks of the Berwick Formation (Zen, editor, 1983). Inclusions of quartzite blocks and fragments are common throughout the pluton. No age determination has been made on the Dracut but correlations with similar rocks of the Exeter Pluton suggest an age of approximately of 470 Ma (Fig. 3.).

Note: This ends Part I. of this roadlog. Stops 3 through 6 (Part II.) will follow a new mileage chart starting from 0.0.

Mileage: Part II.

0.0 Leave George Brox, Inc. quarry and proceed toward Route 110.

0.3 With caution, turn left onto Route 110 and proceed toward Methuen. Be particularly careful crossing traffic lanes as the traffic moves here at rather high speeds.

3.2 Bear left and enter Interstate I-93 going south toward Boston.

3.7 Merrimack River. As you drive over the bridge you're crossing the Clinton-Newbury Fault again. Leaving behind the Merrimack Trough, you will re-enter the Nashoba Block, here underlain by Andover Granite.

6.7 Take Exit 44A from I-93 and proceed on I-495 North toward Lawrence. Roadcuts along I-495 are of Andover Granite.

9.8 Take Exit 42A from I-495 to Route 114 going east.

11.2 Turn left on Route 133 East going toward Georgetown.

11.4 Make another left turn. Route 133 E joins Route 125 E here.

14.1 Lawrence airport. Odometer check.

14.3 Keep right at this Y intersection and follow 133 E for about 7 miles in direction of Gergetown. Route 125 E leaves Route 133 to your left.

21.3 Intersection of Routes 97 and 133 in Georgetown. Continue on Route 133 East.

21.5 Route 133 East turns right; stay on Route 133 E and follow signs in the direction of Route 1.

23.2 Overpass above I-95. About a mile from here, and until you cross Parker River, you will be within the stratified rocks of the Newbury Inlier (Shride, 1976a, 1976b)

26.5 At traffic lights, turn left (N) onto Route 1.

28.9 Turn right (E) on Central Street and almost immediately pull off road and park cars on the shoulder.

STOP #3: NEWBURY INLIER - PORPHYRITIC ANDESITE.

This stop is the same as Stop #2 of Shride, 1976a. On both sides of the street are, what Shride then described as "intercalated flows and water-laid ash-fall (?) tuffs of the porphyritic andesite" (member 7, Shride, 1976a). Here the top of each flow can be recognized by the presence of a sharp transition between vesiculated zones and more massive volcanics of the overlying flows. The same transition also manifests itself by a sudden change in the amount and types of phenocrysts. Fossils found within the Newbury Inlier yield stratigraphic ages of Silurian, possibly through the lowermost Devonian (Shride, 1976b). Four samples, one from each of the major flows from this locality, were analyzed for major and trace elements. Major element data show that the rocks are high alumina basalts, with a narrow range of SiO_2 (near 52%) and a somewhat larger variation for Al_2O_3 (between 17 and 20%). When combined with samples from other locations within the Newbury Inlier, the overall observed variation is like the range of a typical andesitic suite of magmatic arcs. Trace element abundances and various discrimination diagrams further support that the andesites formed in a subduction zone environment. The close association of the basaltic andesites with rhyolites and microgranites within the Newbury Inlier suggests two separate, but contemporaneous, magma systems. Magma mixing between these two respective end-members may have produced a large

array of intermediate rocks such as the Sharpners Pond Quartz Diorite, which will be seen at the next stop.

Turn around, please use EXTREME CAUTION, and prepare to enter Route 1 going north.

29.1 Turn right (N) onto Route 1.

29.4 Newbury town line.

30.5 Parker River bridge. Parker River here follows the Parker River Fault, which separates the Newbury Inlier from the Sharpners Pond Quartz Diorite of the Nashoba Block. Exposures north of the bridge are of pink granite (Sgr of Zen, editor, 1983), and quartz diorite of the Sharpners Pond.

32.8 Turn left on Middle Road, following the sign indicating direction toward Byfield. Stop almost immediately near outcrops on both sides of the road.

<u>STOP #4: SHARPNERS POND QUARTZ DIORITE.</u>
Exposures of the Sharpners Pond Quartz Diorite along both sides of Middle Road show extreme variability in mineralogy, texture, and structure, quite typical for rocks of this general area. Note the complex brecciation and "pillowing" of the more mafic types within the granitic matrix. Our geochemical study shows that the granitic rocks and the mafic to intermediate rocks are genetically unrelated and, in addition, there is some evidence that this granitic component is virtually identical to the granitic masses mapped as pinkish biotite granite near Byfield (Sgr of Zen, editor, 1983). It seems likely that the granitic rocks represent an anatectic melt which formed in response to higher temperatures caused by the intrusions of the more mafic magmas. Such a contemporaneous two-magma system might have interacted in a variety of ways resulting in structures similar to the ones shown by these exposures.

The Sharpners Pond Pluton is a body about 60 mi^2 large, found largely between the towns of Wilmington and Newbury. Castle (1965) identified within the pluton three basic petrographic phases: hornblende diorite, hornblende-biotite tonalite, and biotite tonalite which tend to predominate in a given region, but which show gradational transition from one type to another. Hornblende diorite is almost certainly a cumulate rock consisting of approx. equal amounts of plagioclase (andesine to sodic labradorite) and hornblende (with occasional clinopyroxene cores), and minor amounts (0 to 10%) of biotite. Sphene is a characteristic accessory seen in almost all hand specimens. All other phases of the Sharpners Pond have the same mineralogy (with the exception of minor alkali feldspar) but in varied proportions.

Continue (W) on Middle Road.

33.2 Take sharp turn (N) on unmarked road (Highfield Road).

33.9 Turn left (W) on Scotland Road. More exposures of Sharpners Pond.

35.9 Pass under I-95.

36.2 Make right turn on the side road, and immediately park alongside the road. Walk back few hundred feet toward the roadcuts along the ramp to I-95 South. NOTE: A similar set of outcrops can also be seen along the ramp to I-95 going north.

STOP #5: SHARPNERS POND QUARTZ DIORITE.
These exposures further accentuate the features observed at the previous stop. Well developed magmatic "pillowing", magmatic brecciation and cementing, several stages of subsequent mafic magma intrusions of different types, and magma mixing can all be seen here. Such complex features are not necessarily developed everywhere. As a matter of fact, a short distance west of here and further toward the SW, the rocks are much more uniform in their appearance and more homogeneous in their mineralogy. Age determination on two samples from localities nearby, by $^{206}Pb/^{207}Pb$ on zircons, gave concordant ages of 430+/-5 Ma (Zartman and Naylor, 1984). This age is nearly identical to stratigraphic ages of Upper Silurian obtained for the Newbury Inlier. Considering the additional fact that the Newbury volcanics are also a bimodal suite consisting of rhyolites and andesites, it is then possible to correlate the Newbury inlier as a down faulted block carrying with it the volcanic equivalents of the Sharpners Pond Pluton.

Turn around, drive back under I-95, turn left and enter I-95 going north toward Amesbury.

37.4 Crossing the Clinton-Newbury Fault once again. You are back in the Merrimack Trough. Rock exposures along the highway are of Newburyport Quartz Diorite.

40.4 Merrimack River.

41.1 Take exit from I-95 onto Route 110 (E) in the direction of Salisbury.

43.5 Turn left onto Route 1 (N).

43.6 Set of blinking traffic lights. Bear left on Route 1, and proceed carefully through the intersection.

44.3 Turn right (E) on Gerrish Road, just ahead of railroad overpass.

44.6 Make sharp left turn (N) onto unmarked Seabrook Road.

45.0 Small abandoned quarry on your left. Park in the quarry.

STOP #6: NEWBURYPORT COMPLEX.
This stop is in the same place as STOP #2. of Shride (1971). The Newburyport complex forms a composite pluton which consists of somewhat older tonalites and granodiorites to the SE and of younger porphyritic granites to the NW.

Samples taken from this quarry, which lies within the nonporphyritic tonalites and granodiorites, yielded a $^{207}Pb/^{206}Pb$ zircon age of 466 Ma (Zartman and Naylor, 1984). The porphyritic granite by the same technique gave an age of 437 Ma. The 466 Ma age is remarkably similar to ages obtained for the Exeter and Appledore diorites (Fig. 3) suggesting the presence of a Middle Ordovician magmatic province within the Merrimack Trough. Rocks exposed at this locality are medium grained and are, according to Shride (1971), in the middle of a compositional range defined by mineralogical variations of mafics, feldspars, and quartz. The mafic minerals (near 20%) include biotite and hornblende; sphene is a dominant accessory. A characteristic feature of the Newburyport observed by Shride (1971) is a positive correlation between the frequency of ovoid dioritic inclusions and the more mafic appearance of the host rock. This correlation may be interpreted as evidence for a two-magma system within the Newburyport magma chamber and an incomplete magma mixing between the silicic and more mafic magma type.

After a U-turn, retrace the directions back to Route 1.

45.4 Turn right (W) on Gerrish Road.

45.8 Turn right (N) on Route 1 (Lafayette Road) and proceed under the railroad bridge.

46.6 Traffic lights. Bear left and follow signs in the direction of I-95 going to New Hampshire.

47.5 Continue straight, rejoin I-95 North, and proceed toward Hampton - Exeter Toll Booth exit from I-95 in New Hampshire.

Note: This ends Part II. of this field trip. The upcomming Part III. presents road log to Stops 7 and 7A with fresh start from 0.0 beginning at Hampton - Exeter Toll Both exit off I-95.

Mileage: Part III.

0.0 Tollgate ($0.25 toll). Follow exit ramp to Route 51/101W.

3.6 Traffic lights.

4.0 Pass exit to Route 108 (to Exeter and Newmarket, N.H.)

5.4 Pass exit to Route 85 (to Newfields). Large road cut along Route 101 on your left (S) just beyond the undepass beneath Route 85 exposes contact metamorphosed Kittery Formation. Possible stop (Stop 7A) on the return to I-95.

6.2 Pull off the highway and park near the top of this small hill. 0Exposures of the Exeter Diorite occur on both sides of the highway over the next half mile either as 50 to 100 m long glacially-smoothed pavements or blasted joint surfaces several meters high. The safest crops are on the north (right) side of the highway.

STOP #7: EXETER PLUTON.

The diorite here is fairly typical of the types seen within the Exeter Pluton over its entire 32 by 7 km northeast trending body. It extends from the southernmost exposures, seen here near the town of Exeter, toward Rollingsford, NH. Farther west are exposures of gabbro, that contain plagioclase, orthopyroxene, biotite, minor olivine, and secondary uralitic amphibole. To the northeast, rocks of the Exeter Pluton become gradually more felsic, but never granitic. Over much of its extent, it is typically "salt and pepper", medium gray, medium- to coarse-grained, unfoliated diorite and quartz diorite. Frequent aplite dikes, some pegmatitic, parallel joints, and are sometimes offset by minor faults. Xenoliths of the surrounding Kittery and Eliot Formations are common, particularly near the margins and in areas interpreted to represent the original roof zone (better illustrated in the center of the pluton in Durham). The xenoliths commonly show variable degrees of digestion and often emphasize the carbonate content by the development of epidote, diopside, and occasional grossularite as elongate pods (concretions?). More pelitic inclusions commonly contain coarse, anhedral, poikiloblastic biotite, and small hypersthene granules.

The diorite here contains subequal amounts of plagioclase (3-5 mm subhedra, An35-50), biotite, pyroxenes, amphiboles, quartz (10-15%), and minor magnetite. Comagmatic mafic clots usually 15 - 20 cm long contain abundant biotite, pyroxenes, amphibole, and minor plagioclase. In most thin sections of the Exeter, plagioclase subhedra are normally zoned, very strongly near the contacts with the Merrimack Group. Hornblende is largely altered to chlorite, pyroxene (both clinopyroxene and orthopyroxene occur) remains unaltered, and biotite is slightly chloritized and commonly displays sagenitic rutile. Quartz is invariably interstitial; microcline is occasionally present. Little chemical work has been published on the Exeter Pluton. Preliminary work indicates that the varied body averages 56% silica. Gaudette and others (1984) reported a Rb/Sr whole rock age of 473+/-37 Ma (0.7053+/-0.0005) for the Exeter (Fig.3.), which is consistent with the nearby Newburyport Quartz Diorite, and smaller diorite bodies near Portsmouth and on the Isles of Shoals (Appledore Island data on Fig.3.). The age of these bodies constrains the age of the Merrimack Group in coastal New Hampshire as pre-Middle Ordovician. Other evidence suggests a Late Proterozoic age (Bothner and others, 1984).

Proceed in the same direction toward the next intersection.

7.0 Traffic lights, turn left towards Exeter (a U-turn if possible) and retrace route to I-95.

8.2 STOP 7A: CONTACT METAMORPHOSED KITTERY FORMATION. (Optional stop). If time and interest remains, we can make a short stop at the large outcrop of the contact metamorphosed Ordovician to Precambrian(?) Kittery Formation exposed on the west side of the Route 85 bridge. The outcrop contains brittle, purplish-brown quartzite and pelitic hornfels, and is less than 1 km from the contact with the Exeter Pluton. The actual contact is not exposed here. The quartzite occurs in beds 30 to 50 cm thick intercalated with 5 - 10 cm pelitic hornfels (originally phyllite) layers. Occasional calc-silicate bands occur within the quartzite. The surface facing the highway is strongly slickensided. On the top of the outcrop, porphyroblasts of probable cordierite (now retrograded to white mica), form noticeable knots 1-10 mm across in some pelitic layers. Elsewhere at the contact, hypersthene is developed within the pelitic portions of the Kittery. They likewise show retrograde alteration. It is therefore likely that the Kittery, regionally metamorphosed before the emplacement of the Exeter, was contact metamorphosed some 473 Ma ago, and then mildly metamorphosed during the Acadian or Alleghanian events.

Continue back to I-95. Proceed north to Lewiston, Maine

End of Part III. and end of field trip.

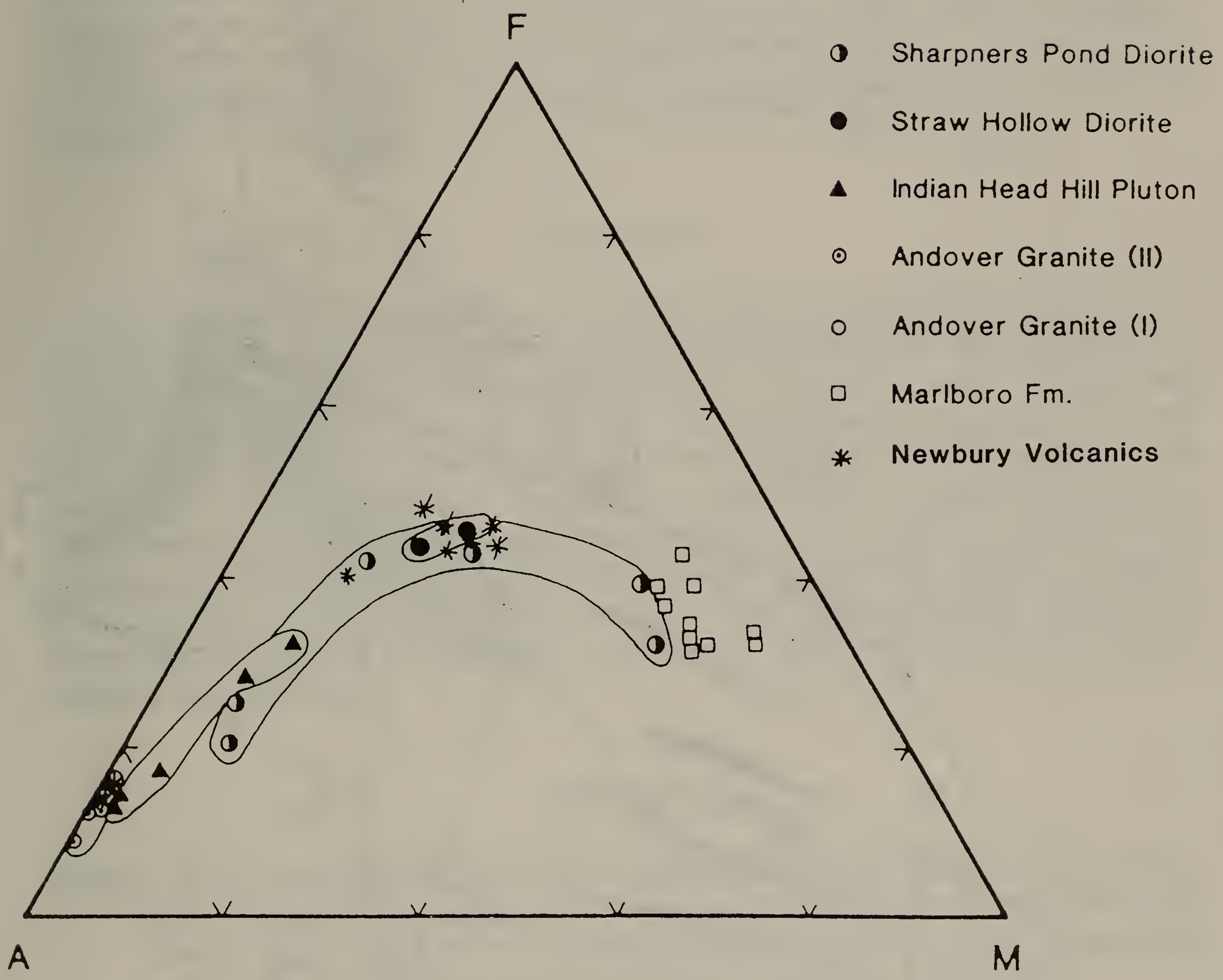
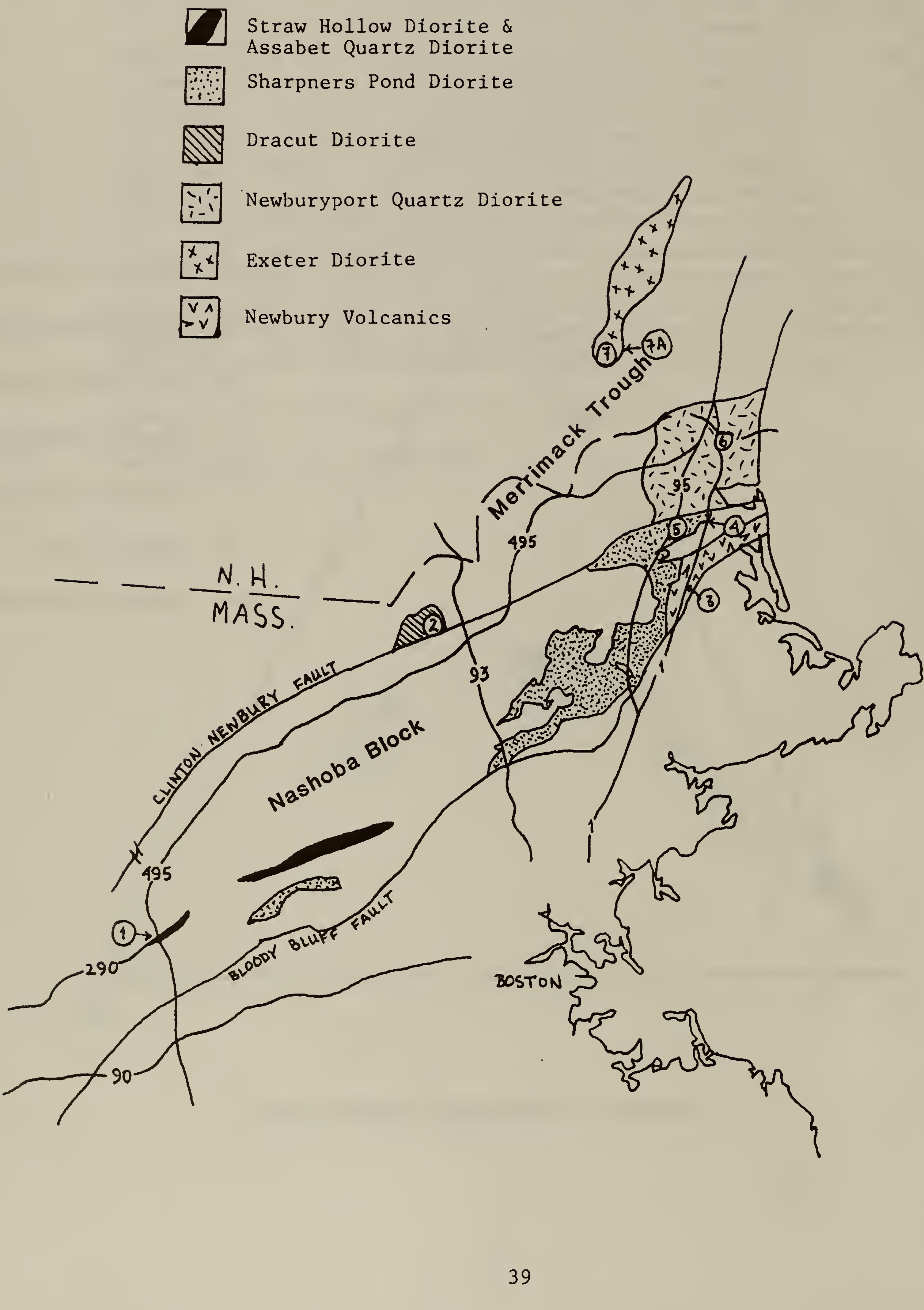


FIGURE 1. AFM Diagram Nashoba Block

FIGURE 2. Stop Locations



MERRIMACK TROUGH

Merrimack Trough has only recently been recognized as a seperate terrane (Lyons and others, 1982). More work needs to be done but many recent contributions indicate that this region is indeed a separate terrane (see for example: Bothner and others, 1984; Gaudette and others, 1984). One of the rather interesting and tantalizing suggestions is that the Merrimack Trough is in fact an old Precambrian Block which suffered insignificant deformation during the Acadian Orogeny yet contains calc-alkalic rocks of the type indicative of continental margins associated with subduction zones. If the ages of these intrusions can be used to time the presumed subduction zone then the subduction lasted from around 470 Ma ago to about 400 Ma ago.

CONCLUSIONS

At present not enough data exists relevant to the significance of the mid-Paleozoic calc-alkalic magmatism in northeastern Massachusetts and southeastern New Hampshire. Several M.Sc. theses currently underway at Boston College may perhaps provide some of the key data, and if these are available by the time this field trip is being conducted, they will be informally shared with all the participants. In conclusion, we would like to suggest that the mid-Paleozoic calc-alkalic suite is indicative of a long lasting (appr. 70 Ma) subduction zone which might have ultimately led to a continental collision, the Acadian Orogeny. Position of the arc axis probably was not at the same place throughout the entire time period but most likely it migrated from place to place leaving behind plutonic bodies of discrete ages.

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ITINERARY

<u>Mileage: Part I.</u>

0.0 Assembly point and Stop #1. Meet at junction of Interstates I-290 and I-495 just north of Marlborough, Mass. at 10 a.m. Meet on S.E. side of this large interchange. If coming north on I-495, take exit 25A "to 85, Marlborough", park on shoulder of ramp just before it merges with I-290 by silver colored electric box. If coming east on I-290, continue over I-495 and park just east of where the ramp from I-495N, joins I-290 extension "to 85, Marlborough". We will leave this exposure at <u>11 a.m.</u>

<u>STOP #1: STRAW HOLLOW DIORITE or ASSABET QUARTZ DIORITE</u>
The large exposures at this cloverleaf are complex and show a wide variety of features. Our principal purpose in stopping here is to examine the Straw Hollow Diorite or Assabet Quartz Diorite along the north side of the interchange. The diorite here is characteristic of the smaller bodies of Silurian calc-alkaline plutonic rocks in the Nashoba Block. Both foliated and non-or more weakly foliated varieties of hornblende-quartz diorite are present here. They have been intruded by garnet bearing pegmatites associated with the younger phase of the Andover Granite (approximately 415 m.y., Hill et al., 1984). Sillimanite schists of the Nashoba Fm. are exposed along the south side of the cut.

Blastomylonites believed to be associated with the Assabet River fault zone are also exposed here and will be briefly examined as an example of the type of deformation common along some of the larger shear zones in the Nashoba Block. If time permits, a relatively late brittle shear zone cutting the Straw Hollow, with carbonate mineralization, will be visited to provide contrast with the earlier ductile deformation.

To proceed to Stop 2 we will need to turn around and head north on I-495.Continue straight ahead on "to 85" east.

1.0 At traffic light, Fitchburg Street, exit to the right.

1.1 By entrance to Assabet Valley Regional Vocational High School, turn around and retrace route to traffic light.

1.2 Turn left (west) on "to 495-290".

2.2 Take Exit 26B, I-495 north toward Lowell.

2.8 Merge with I-495 north.

3.0 Cross Assabet River.

3.8 Excellent exposures of Nashoba Fm. gneisses and migiratites for next several miles, continue north toward Lowell.

25.8 Rest area on right.

26.7 Leave I-495 at Exit #36, "Lowell Connector". Exit to right.

27.4 Turn right, follow Exit 36, Lowell Connector.

27.5 Bear left, follow Exit 36, Lowell Connector.

28.1 Continue straight on Lowell Connector.

28.8 Continue toward Lowell Center.

29.9 Road ends, turn left toward Lowell Center.

30.1 Continue straight past Courthouse on right.

30.2 Bear right, follow one way.

30.4 Bear left, continue on one way.

30.45 At light, continue straight, now on Central St.

30.6 At light, Y intersection, bear right on Prescott St.

30.7 At light, turn right on E. Merrimack St.

30.8 Federal Building on left.

31.0 Continue straight.

31.1 At light, turn left on routes 38 and 110.

31.3 Cross Merrimack River and bear right on Rte. 110 east.

31.4 Continue on Rte. 110 east toward Lawrence.

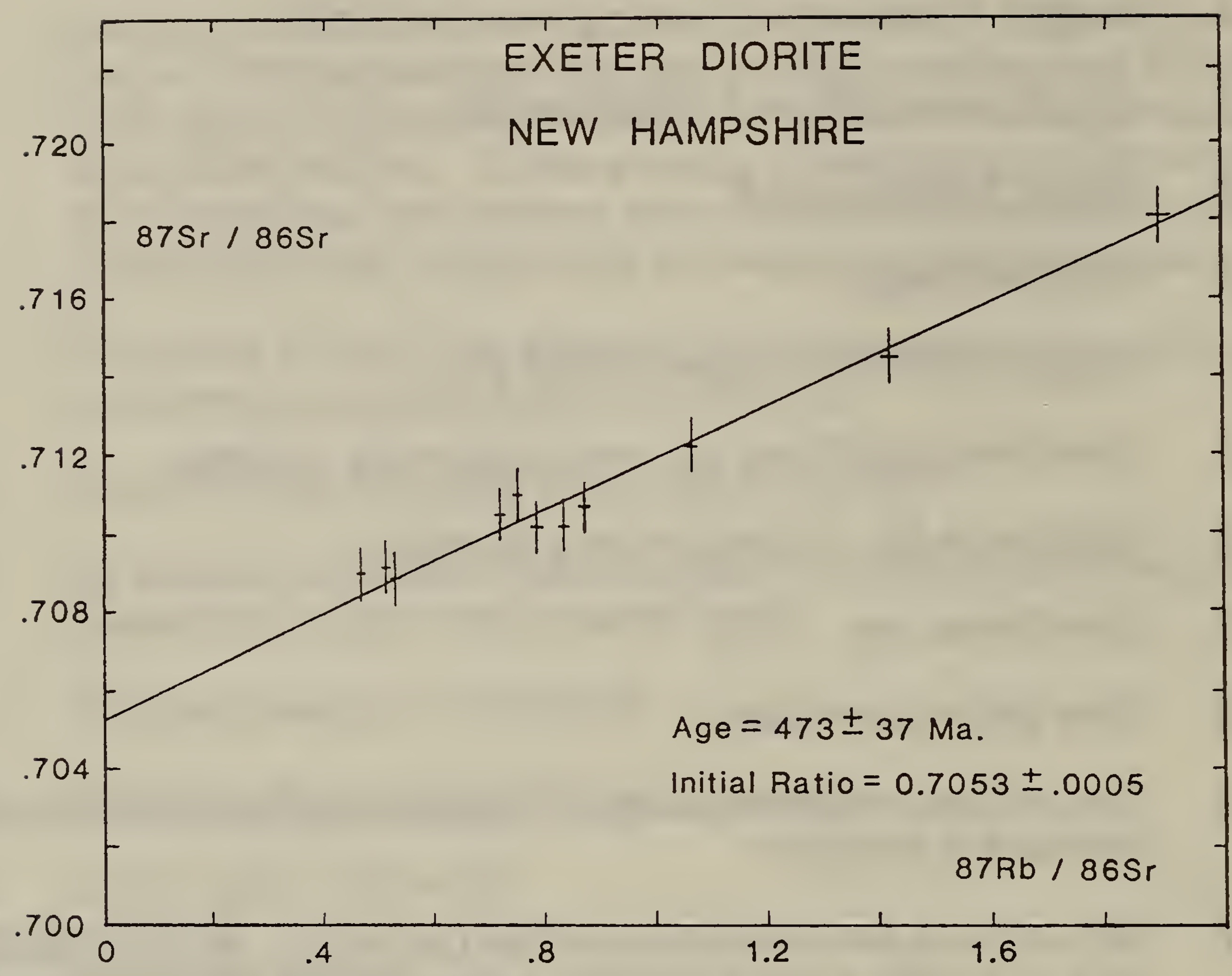
32.2 Dracut town line.

34.9 Cross under power lines.

35.3 Turn left onto side road at sign for George Brox, Inc., proceed up hill. WATCH FOR TRUCKS!

35.7 Park in lot on left at entrance to Brox Quarry,Inc. WATCH FOR TRUCKS-THIS IS A VERY ACTIVE QUARRY. At this locality we will consolidate to as a few vehicles as possible and proceed driving into the quarry. Prepare to stop at the gatehouse and obtain a clearance or permition to enter the quarry. For this trip we are promissed an entrance for all participants. We will proceed driving to the lower level and park near an area which does not interfere with the quarry operation.

STOP #2: DRACUT DIORITE.
In this quarry virtually every petrographic type known to occur within the Dracut stock can be found. Main rock type is, however, a massive, medium to coarse grained unfoliated norite. This body crops out over an area of about 27 mi^2, intruding quartzites of the Berwick Formation (Zen, editor, 1983). Inclusions of quartzite blocks and fragments are common throughout the pluton. There is no age determination but correlations with similar rocks of the Exeter Pluton suggests a date in the vicinity of 470 Ma (Fig.3). Petrographic studies done by Dennen (1942) show that the mineralogy is highly variable from noritic diorite to cumulate ultramafics consisting of 55% forsteritic olivine, 20% enstatitic orthopyroxene, 5% each bytownite and clinopyroxene, and 15% of intercumulus biotite and hornblende. Predominant rock type is norite with subequal amounts of plagioclase (An71-An37) and mafics (cpx, opx and olivine) with minor biotite and hornblende. Dennen (1942,1943) describes a gradation from more felsic varieties near the



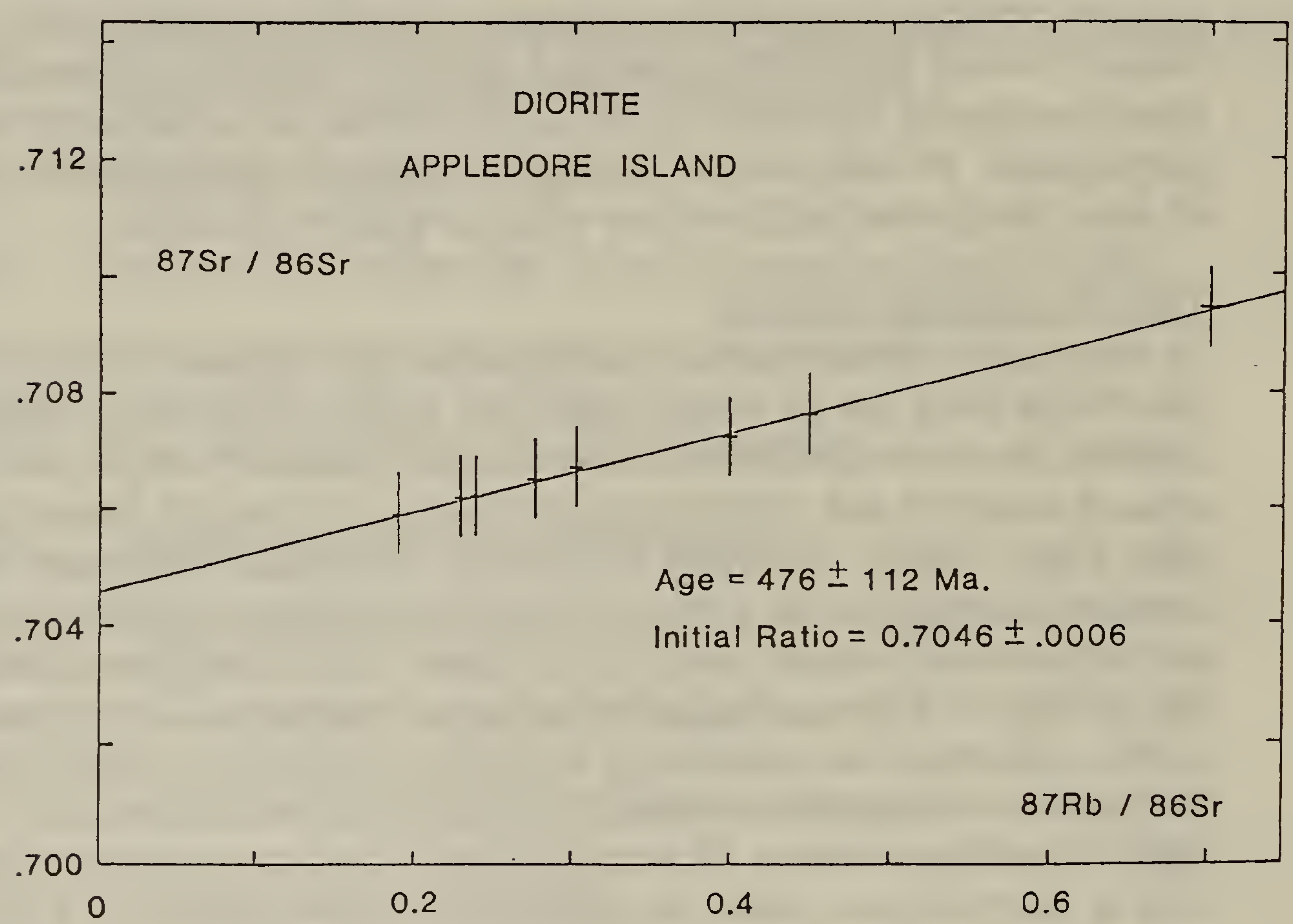


FIGURE 3 Rb/Sr Isochrons for Exeter and Appledore Island Plutons
Rb Decay Constant = 1.42×10^{-11} Years.

contacts to more mafic norites in the center noting that this zonation is likely due to magmatic fractionation. A curious occurrence of magmatic sulphides which can reach locally quite high proportions (over 50%), led to a prospecting effort for nickel - hence Nickel Mine Hill, the location of this quarry.

Note: This ends Part I. of this roadlog. Stops 3 through 6 (Part II.) will follow a new mileage chart starting from 0.0.

Mileage: Part II.

0.0 Leave George Brox, Inc. quarry and proceed toward Route 110.

0.3 With caution, turn left onto Route 110 and proceed toward Methuen. Be particularly careful crossing traffic lanes as the traffic moves here at rather high speeds.

3.2 Bear left and enter Interstate I-95 going south toward Boston.

3.7 Merrimack River. As you drive over the bridge you're crossing the Clinton-Newbury Fault again. Leaving behind the Merrimack Trough, you will re-enter the Nashoba Block, here underlain by Andover Granite.

6.7 Take Exit 44A and proceed on I-495 North toward Lawrence. Roadcuts along I-495 are of Andover Granite.

9.8 Take Exit 42A from I-495 to Route 114 going east.

11.2 Turn left on Route 133 East going toward Georgetown.

11.4 Make another left turn. Route 133 E joins Route 125 E here.

14.1 Lawrence airport.

14.3 Keep right at this Y fork and follow 133 E for about 7 miles in direction of Gergetown. Route 125 E leaves Route 133 to your left.

21.3 Intersection of Routes 97 and 133 in Georgetown. Continue on Route 133 East.

21.5 Route 133 East turns right; stay on Route 133 E and follow signs in the direction of Route 1.

23.2 Overpass above I-95. About a mile from here, and until you cross Parker River, you will be within the stratified rocks of the Newbury Inlier (Shride, 1976a,1976b)

26.5 At traffic lights, turn left (N) onto Route 1.

28.9 Turn right (E) on Central Street and almost immediately pull off road and park cars on the shoulder.

STOP #3: NEWBURY INLIER - PORPHYRITIC ANDESITE: This stop is the same as Stop #2 of Shride, 1976a. On both side of the street are exposures of "intercalated flows and water-laid ash-fall (?) tuffs of the porphyritic andesite (member 7)" (Shride,1976a). Fossils from a locality near here and other places within the Inlier yield stratigraphic ages of uppermost Silurian, possibly through the lowermost Devonian (Shride, 1976b). Four samples, one from each of the flows, were analyzed for major and trace elements. Major element data show that these are high alumina basalts: SiO_2 near 52% and Al_2O_3 between 17 and 20%. Trace elements and various discrimination diagrams suggest a subduction zone related magmatism. Other analyzed samples from the Newbury Inlier, however, are more like the typical andesites of magmatic arcs. Top of each flow is here readily recognizable by the presence of a vesicular band. Also note the differences in the phenocrystic content between the flows.

Close association of the basaltic andesites with rhyolites, and microgranites within the Newbury Inlier, is suggestive of two separate, but contemporaneous magma systems. Magma mixing between the two respective end-members could produce a large array of intermediate rocks such as the one of Sharpners Pond Quartz Diorite, which will be seen at the next stop.

Turn around, please use EXTREME CAUTION, and prepare to enter Route 1 going north.

29.1 Turn right (N) onto Route 1.

29.4 Newbury town line.

30.5 Parker River bridge. Parker River follows here the Parker River Fault, which separates the Newbury Inlier from Sharpners Pond Quartz Diorite of the Nashoba Block. Exposures north of the bridge are of pink granite (Sgr of Zen, editor, 1983), and quartz diorite of Sharpners Pond.

32.8 Turn left on Middle Road by following the sign indicating direction toward Byfield. Stop almost immediately after the turn alongside the road.

STOP #4: SHARPNERS POND QUARTZ DIORITE.

Examine exposures of Sharpners Pond Quartz Diorite along both sides of Middle Road just as it approaches Route 1, and observe the extreme variability in mineralogy, texture, and structure. These exposures are rather typical of rocks in this general area. Note the complex brecciation and "pillowing" of more mafic types within a granitic matrix. There is some evidence that the granitic component is virtually the same as the granitic masses mapped as pinkish biotite granite near Byfield (Zen, editor, 1983). Geochemical study shows that the granite and the mafic to intermediate rocks are genetically unrelated. Perhaps the granitic rocks represent anatectic melts which formed in response to raised temperatures due to intrusions of more mafic magmas. Such two magma system would interact in a variety of ways resulting in structures shown on these exposures.

Sharpners Pond Pluton forms about 60 mi^2 large body found between Wilmington and Newbury. Castle (1965) identified within the pluton three basic petrographic phases: hornblende diorite, hornblende-biotite tonalite, and biotite tonalite which tend to predominate in a given region but which show graditional transition from one type to another. Hornblende diorite is almost certainly a cumulate rock consisting of subequal amounts of plagioclase (andesine to sodic labradorite) and hornblende with occasional clinopyroxene cores, and minor amounts (0 to 10%) of biotite. Sphene is a characteristic accessory seen in almost all hand specimen. All other phases of the Sharpners Pond have the same mineralogy (with the exception of minor alkali feldspar) but in varied proportions.

Continue (W) on Middle Road.

33.2 Take sharp turn (N) on unmarked road (Highfield Road).

33.9 Turn left (W) on Scotland Road. More exposures of Sharpners Pond.

35.9 Pass under I-95.

36.2 Make right turn on the side road, and immediately park alongside the road. Walk back few hundred feet toward the roadcuts along the ramp to I-95 South. NOTE: A similar set of outcrops can also be seen along the ramp to I-95 going north.

STOP *5: SHARPNERS POND QUARTZ DIORITE.
These exposures further accentuate the same and other similar features as observed at the previous stop. Well developed magmatic "pillowing", magmatic brecciation and cementing, several stages of subsequent mafic magma intrusions and of different types, and magma mixing. Not everywhere we see these complex features, as a matter of fact a short distance from here, and further toward SW the rocks are much more uniform in their appearance and more homogeneous in their mineralogy.
Age determination on two samples from nearby localities by $^{206}Pb/^{207}Pb$ on zircons gave concordant ages of 430+/-5 Ma. This age is nearly identical to stratigraphic ages of Upper Silurian obtained for the Newbury Inlier. Considering the additional fact that the Newbury volcanics are also a bimodal suite consisting of rhyolites and andesites, it is possible then to correlate the Newbury inlier as a down faulted block carrying with it volcanic equivalents of the plutonic rocks of the Sharpners Pond Pluton.

Turn around, drive back under I-95, and enter I-95 going north toward Amesbury.

37.4 Crossing the Clinton-Newbury Fault once again. You are back in the Merrimack Trough. Rock exposures along the highway are of Newburyport Quartz Diorite.

40.4 Merrimack River.

41.1 Take exit from I-95 onto Route 110 (E) in the direction of Salisbury.

43.5 Turn left onto Route 1 (N).

43.6 Set of blinking traffic lights. Bear left on Route 1, and proceed carefully through the intersection.

44.3 Turn right (E) on Gerrish Road, just ahead of railroad overpass.

44.6 Make sharp left turn (N) onto unmarked Seabrook Road.

45.0 Small abandoned quarry on your left. Park in the quarry.

STOP *6: NEWBURYPORT COMPLEX.
This stop is in the same place as STOP *2. of Shride (1971).
Newburyport complex forms a composite pluton which consists of somewhat older tonalites and granodiorites to the SE and younger porphyritic granites to the NW. Samples taken from this quarry, which

lies within the nonporphyritic tonalites and granodiorites, yielded $^{207}Pb/^{206}Pb$ zircon age of 466 Ma (Zartman and Naylor, 1984), whereas the porphyritic granite gave by the same technique an age of 437 Ma. The older age is remarkably similar to ages obtained for the Exeter and Appledore diorites (Fig. 3) suggesting a presence of Middle Ordovician magmatic province within the Merrimack trough.
Rocks exposed at this locality are medium grained and are, according to Shride (1971), in the middle of the range defined by mineralogical variations of mafics, feldspars and quartz. The mafic minerals (near 20%), include biotite and hornblende, sphene is a dominant accessory. Characteristic feature observed by Shride (1971) is a positive correlation between the frequency of ovoid dioritic inclusions and the more mafic appearance of the host rock which may be interpreted as an evidence for mixing between two types of magmas..

After a U-turn, retrace the directions back to Route 1.

45.4 Turn right (W) on Gerrish Road.

45.8 Turn right (N) on Route 1 (Lafayette Road) and proceed under the railroad bridge.

46.6 Traffic lights. Bear left and follow signs in the direction of I-95 going to New Hampshire.

47.5 Continue straight, rejoin I-95 North, and <u>proceed toward Hampton - Exeter Toll Booth exit from I-95 in New Hampshire.</u>

<u>Note:</u> This ends <u>Part II.</u> of this field trip. The upcomming <u>Part III.</u> presents road log to Stops 7 and 7A with fresh start from 0.0 beginning at Hampton - Exeter Toll Both exit off I-95.

<u>Mileage: Part III.</u>

0.0 Toolgate ($0.25 toll). Follow exit ramp to Route 51/101W.

3.6 Traffic lights.

4.0 Pass exit to Route 108 (to Exeter and Newmarket, N.H.)

5.4 Pass exit to Route 85 (to Newfields). Large road cut along Route 101 on your left (S) just beyond the undepass beneath Route 85 exposes contact metamorphosed Kittery Formation. Possible stop (Stop 7A) on the return to I-95.

6.2 Pull off the highway and park near the top of this small hill. 0Exposures of the Exeter Diorite occur on both sides of the highway over the next half mile either as 50 to 100 m long glacially-smoothed pavements or blasted joint surfaces several meters high. The safest crops are on the north (R) side of the highway.

STOP #7: EXETER PLUTON.

The diorite here is fairly typical of the types seen within the Exeter Pluton over its entire 32 by 7 km northeast - trending body. It extends from the southern exposures, seen here near the town of Exeter, toward Rollingsford, NH. Farther west are exposures of gabbro, that contain plagioclase, orthopyroxene, biotite, minor olivine, and secondary uralitic amphibole. To the northeast, rocks of the Exeter Pluton become gradually more felsic, but never granitic. Over much of its extent, it is typically "salt and pepper", medium gray, medium- to coarse-grained, and unfoliated diorite and quartz diorite. Frequent aplite dikes, some pegmatitic, parallel joints, and are sometimes offset by minor faults. Xenoliths of the surrounding Kittery and Eliot Formations are common, particularly near the margins and in areas interpreted to represent the original roof zone (better illustrated in the center of the pluton in Durham). The xenolith commonly show variable degrees of digestion and often emphasize the carbonate content by the development of epidote, diopside, and occasional grossularite as elongate pods (concretions?). More pelitic inclusions commonly contain coarse, anhedral, poikiloblastic biotite, and small hypersthene granules.
The diorite here contains subequal amounts of plagioclase (3-5 mm subhedra, An35-50), biotite, pyroxenes, amphiboles, quartz (10-15%), and minor magnetite. Comagmatic mafic clots usually 15 - 20 cm long contain abundant biotite, pyroxenes, amphibole, and minor plagioclase. In most thin sections of the Exeter, plagioclase subhedra are normally zoned, very strongly near the contacts with the Merrimack Group. Hornblende is largely altered to chlorite, pyroxene (both clinopyroxene and orthopyroxene occur) remains unaltered, and biotite is slightly chloritized and commonly displays sagenitic rutile. Quartz is invariably interstitial; microcline is occasionally present. Little chemical work has been published on the Exeter Pluton. Preliminary work indicates that the varied body averages 56% silica. Gaudette and others (1984), reported a Rb/Sr whole rock age of 473+/-37 Ma (0.7053+/-0.0005) for the Exeter (Fig.3.), which is consistent with the nearby Newburyport Quartz Diorite, and smaller diorite bodies near Portsmouth and on the Isles of Shoals (Appledore Island data on Fig.3.). The age of these bodies constrains the age of the Merrimack Group in coastal New Hampshire as

pre-Early Ordovician. Other evidence suggests a Late Proterozoic age (Bothner and others, 1984).

Proceed in the same direction toward the next intersection.

7.0 Traffic lights, turn left towards Exeter (a U-turn if possible) and retrace route to I-95.

8.2 <u>STOP 7A: CONTACT METAMORPHOSED KITTERY FORMATION.</u> (Optional stop). If time and interest remains, we can make short stop at the large outcrop of the contact matamorphosed Ordovician to Precambrian(?) Kittery Formation exposed on the west side of the Route 85 bridge. The outcrop of brittle, purplish-brown quartzite and pelitic hornfels within less than 1 km of the contact with the Exeter Pluton. The actual contact is not exposed here. The quartzite occurs in beds 30 to 50 cm thick intercalated with 5 - 10 cm pelitic hornfels (originally phyllite) layers. Occasional calc-silicate bands occur within the quartzite. The surface facing the highway is strongly slickesided. On the top of the outcrop, porphyroblasts of probable cordierite (now retrograded to white mica), which form noticeable knots 1-10 mm across, occur in some peliticlayers. Elsewhere at the contact, hypersthene is developed within the pelitic portions of the Kittery. They likewise show retrograde alteration. It is therefore likely that the Kittery, regionally metamorphosed before the emplacement of the Exeter, was contact metamorphosed some 473 Ma ago, and then mildly metamorphosed during the Acadian or Alleghanian events.

14.0 Hampton Toll Gate, I-95. Proceed north to Lewiston, Maine

End of <u>Part III.</u> and end of field trip.

GEOLOGICAL COMPARISONS ACROSS THE NORUMBEGA FAULT ZONE, SOUTHWESTERN MAINE

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INTRODUCTION

The Norumbega Fault Zone is a genetically related series of faults extending from New Brunswick nearly to Long Island Sound. The zone is widest in southwestern Maine, the area of this field trip (Figure 1). The most significant of the faults within the zone in southwestern Maine appears to be the Flying Point Fault (Fig.1). South of Scarboro, the fault zone is represented by the Nonesuch River Fault with which the Flying Point Fault merges. The Nonesuch River Fault continues into New Hampshire where it apparently is the same break as the Campbell Hill Fault of Lyons, et al. (1982).

In southwestern Maine the Norumbega Fault Zone forms the boundary between the Central Maine sequence and the Merrimack Group, and between the Central Maine sequence and the Casco Bay Group southwest of Portland. To the northeast, in the eastern Maine area, rocks of Late Ordovician to possibly Devonian age (the Vassalboro, Bucksport, and Flume Ridge Formations) have been mapped on either side of the Fault Zone, but have been displaced only a few 10's of kilometers at most. The Fault Zone is essentially locked by plutons of Carboniferous age (Biddeford, Saco, and Lyman; Hussey and Newberg,1978), and may be more significant with respect to basement terranes, as discussed below.

This field trip will focus on geological contrasts and similarities of terranes on either side of the Norumbega - Nonesuch River Fault Zone in southwestern Maine. Stops will be made to examine lithologies and structures of formations of the Merrimack Group, the central Maine sequence, and the Casco Bay Group. Participants will have an opportunity to discuss the significance of radiometric dates which apparently deny correlations of similar rock units on either side of the Fault Zone.

STRATIGRAPHY AND STRUCTURE OF SOUTHWESTERN MAINE

Four terranes comprise the bedrock of southwestern Maine: 1) the Rye Terrane consisting of the Rye Formation, 2) the Central Maine Terrane consisting of the Vassalboro, Windham, and Waterville Formations and formations of the Shapleigh Group; 3) The Casco Bay Terrane consisting of the Cushing, Cape Elizabeth, Spring Point, Diamond Island, Scarboro, Spurwink, Jewell and Macworth Formations; and 4) the Merrimack Terrane consisting of the Kittery, Eliot, and Berwick Formations.

RYE TERRANE. The very southwestern tip of Maine is underlain by the Rye Formation which extends southward in New Hampshire to the Seabrook area where it plunges to the southwest beneath the Kittery Formation of the Merrimack Group. It is restricted to the southeast side of the Norumbega Fault. The Rye Formation consists principally of regionally mylonitized metasedimentary rocks (mostly metashales and metasiltstones, in part calcareous) which have been migmatized and pegmatite-injected to varying degrees (Hussey, 1980; Carrigan, 1984a, b, and c). The most heavily migmatized rocks were originally interpreted to be a sequence of felsic metavolcanic rocks, but detailed studies have shown that 1) some of the migmatized rocks have abundant sillimanite and relic staurolite and andalusite, and 2) the felsic stringers occasionally transect compositional layering interpreted to be bedding in the metasediments. Minor lithologies include amphibolite, rusty schist, and impure marble. At the north edge of its outcrop belt the Rye Formation is in contact with the Kittery Formation across an ultramylonite zone 75m or so wide representing deep ductile strike-slip or thrust faulting. Swanson (personal communication, 1986) regards this to be right-lateral strike-slip motion. The Rye Formation is correlated with the Nashoba Formation, and pelitic parts of the Massabesic Gneiss (Hussey, 1985) and is probably late Precambrian in age.

MERRIMACK TERRANE. The Merrimack Terrane is underlain by the Kittery, Eliot, and Berwick Formations, an apparently conformable sequence of calcareous and feldspathic metaturbidites. These formations are restricted to the southeast side of the Norumbega Fault.

The Kittery Formation consists of thin to thick, variably bedded calcareous and feldspathic metawacke, occasionally with coarse sand sized clasts of quartz, feldspar,and dark rock fragments in the bases of thick graded beds. Minor sedimentary structures in addition to graded bedding include cross-bedding, flame structures, minor channel cut and fill, and parallel laminae (Rickerich, 1983). Rickerich (1983) interprets the environment of deposition of the Kittery Formation to be that of a deep-sea fan. He shows from analysis of direction of inclination of the foresets of crossbeds that the beds of the Kittery Formation were deposited primarily by currents flowing

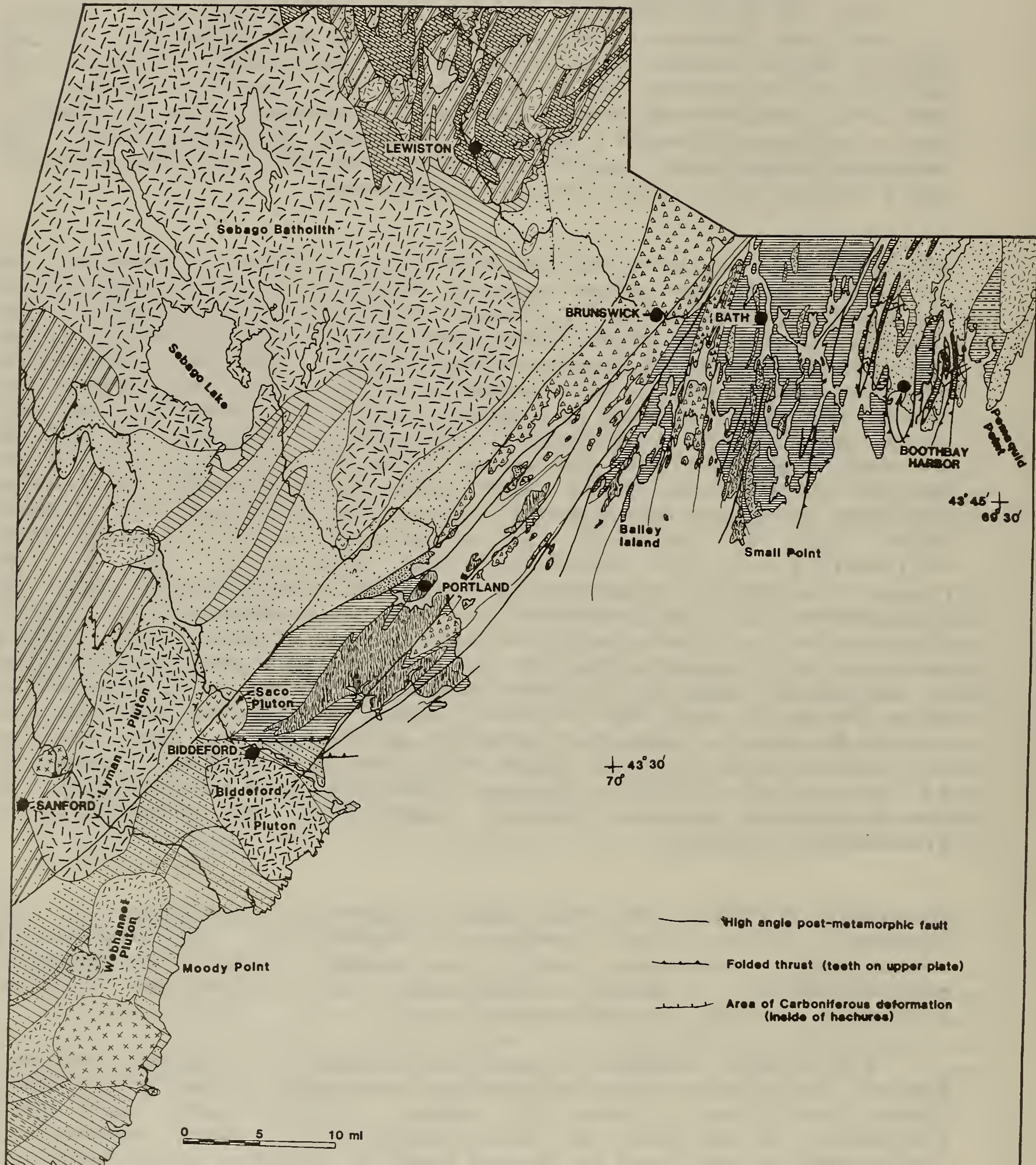

Figure 1. Generalized geologic map of southwestern Maine.

EXPLANATION

INTRUSIVE ROCKS

Mesozoic	Gabbro, alkaline granite,and related intrusives
Carboniferous	2-mica granite
Carboniferous or older	Foliated gabbro-diorite
E. Devonian	2-mica granite
	Foliated granodiorite

STRATIFIED ROCKS

E. Sil.	Rindgemere, Sangerville Fms, Patch Mtn. M.,Sangerville	CENTRAL MAINE SEQUENCE
	Windham, Waterville Fms., Anasagunticook M.,Sangerville Fm.	
L. Ord. to E. Sil.	Vassalboro Fm.	
PreЄ ?	Berwick Fm.	MERRIMACK GROUP
	Eliot Fm.	
	Kittery Fm.	
PreЄ ? to Ord ?	Macworth Fm.	CASCO BAY GROUP
	Jewell,Spurwink,Scarboro, Diamond Island, Spring Point Fms.	
	Cape Elizabeth Fm.	
	Cushing Fm.	
PreЄ	Rye Fm.	

L. Ord. to E. Dev.

Bucksport Fm.

Cross River Fm.

from east to west. This he interprets to be the original paleoslope.

The Eliot formation consists of thin-bedded alternations of calcareous and ankeritic metasiltstone and dark chlorite phyllite, with a basal zone lacking the dark phyllite. The contact with the Kittery Formation is conformable and relatively abrupt.

The Berwick Formation is very similar to the Kittery Formation but is only exposed in areas of higher metamorphic grade. It consists of thin and medium bedded to massive quartz-plagioclase-biotite-actinolite granofels with interbeds of greenish gray calc-silicate granofels which may be locally abundant as at the type locality in Berwick, Me. Diopside and zoisite are common in the northwestern part of the outcrop belt and grossularite is present in pods and metamorphically differentiated veins. Contacts with the Eliot Formation are not exposed but the two formations are considered to be conformable.

The Kittery Formation is affected by three major folding events. The oldest folds are recumbent south to southeast-facing, north to northwest-verging isoclines best seen along the Ogunquit shoreline (Hussey et al., 1984). These have been refolded by upright folds whose axes show frequent plunge reversals, with plunges seldom exceeding 25 degrees. The latest folds are relatively open and northwest verging, and have a strong axial-planar spaced cleavage which in the Ogunquit area is the principal cleavage observed (Hussey et al. 1984). The deformational character of the Eliot and Berwick Formations is not well known because of the poorer exposures of these formations away from the coast; the intervening Calef Member in New Hampshire, however, displays a very strong phyllonitic character.

The age of the Merrimack Group is regarded by Gaudette et al. (1984) to be Late Precambrian to earliest Ordovician on the basis of a Rb/Sr age of 473+73 Ma obtained for the Exeter Pluton which post-tectonically intrudes the Kittery and Eliot Formations, and a 450 Ma zircon age reported by Zartman and Naylor (1984) for the Newburyport Quartz Diorite which intrudes the Kittery Formation in the Newburyport, Massachusetts, area. In addition, Bothner et al. (1984) have interpreted a gradational contact between the Berwick and the Massabesic Gneiss Complex with no metamorphic or significant structural breaks. On the other hand, on a lithologic basis, the Berwick and Kittery Formations are similar to the Vassalboro Formation of the Central Maine sequence suggesting the possibility of a late Ordovician to earliest Silurian age for the Merrimack Group. For a matter of tectonic interpretation as significant as this, additional ages for these plutons, particularly the Exeter, should be obtained by other methods of radiometric dating. If the same age range holds up then a correlation of the Merrimack Group with the Vassalboro will be precluded and the Merrimack sequence must then be interpreted as a totally

independent terrane.

CENTRAL MAINE TERRANE. The formations of the Central Maine Terrane in the general area of the field trip include the Vassalboro, Waterville, Windham, and Sangerville Formations, and the formations of the Shapleigh Group. Only the Vassalboro and Windham Formations will be seen on this trip. The eastern edge of the Vassalboro Formation has been cut by the Norumbega Fault, thus parts of the Central Maine Terrane lie east of the Fault.

The Vassalboro Formation in the area of this field trip consists of quartz-plagioclase-biotite (-hornblende) granofels with or without calc-silicate granofels interbeds. It has been metamorphosed to staurolite and higher grade. East of the Westbrook tongue of the Sebago Batholith, the Vassalboro is extensively migmatized.

The Windham Formation which is correlated with the Waterville Formation on the basis of both lithic similarity and similarity of sequence consists of thin-bedded biotite-muscovite-garnet-quartz-plagioclase schist, and biotite granofels (Thomson, 1985a and b). Staurolite, sillimanite, and kyanite are present at the respective grades of metamorphism. ribbon metalimestone, consisting of thin-bedded, fine-grained, gray marble with thin interbeds of quartzose mica schist forms a 50m thick member in the middle of the Formation. Associated closely with the ribbon metalimestone is calc-silicate granofels and biotite granofels. Calc- silicate minerals present include diopside, grossularite, green amphibole, and calcic plagioclase (Thomson, 1985a and b).

The central Maine sequence has been affected by two major deformations. The earlier produced large-scale recumbent folds (Fl) as described by Osberg (1980) in the Waterville area, and the later produced large-scale upright to slightly overturned folds (F2) that will be seen on this field trip at the Union Falls stop. Fl and F2 folds are a result of the Acadian Orogeny of early Devonian age. F3 folds are very minor structures related to intrusion of the Lyman and Sebago plutons, both of which are Carboniferous in age, hence F3 folds are of Carboniferous age. These rock, which will be seen at Union Falls (Stop 3), were metamorphosed to staurolite grade during the Acadian orogeny, and remetamorphosed in Carboniferous time after emplacement of the Sebago Batholith (Thomson, 1985a and b). This later metamorphism, related spatially to the bottom of the Batholith, produced the kyanite in the South Windham area which will be the focus of our stop at Dundee Falls (Stop 4).

CASCO BAY TERRANE. The Casco Bay Terrane is underlain by the Casco Bay Group consisting of the Cushing, Cape Elizabeth, Spring Point, Diamond Island, Scarboro, Spurwink, Jewell, and Macworth Formations in ascending stratigraphic order. The contacts between formations are conformable 2xcept that between

the Cape Elizabeth and Cushing Formations which may be, at least locally, unconformable.

The Cushing Formation is composed largely of felsic to intermediate metavolcanic rocks, and feldspathic volcanogenic metasedimentary rocks , with lesser mafic metavolcanic rocks, calc-silicate granofels, marble, and sulfidic schist. In the belt from South Portland through the central part of Casco Bay to Harpswell Neck, on the east side of the Norumebega Fault, the Formation shows the most distinctive volcanic character of any part of its outcrop belt. The rocks are very feldspathic, and structures indicate that much of the rock was crystal tuff and volcanic breccia prior to metamorphism. The Formation shows significant facies changes across the strike of the belt, but relatively constant character parallel to strike. Across strike to the east, the upper part of the Formation has a stronger clastic character with abundant calc-silicate granofels and minor sillimanite-bearing feldspathic granofels.

West of the Norumbega Fault, rocks mapped as the Cushing consist predominantly of feldspathic metasedimentary rocks and felsic metavolcanic rocks in part with abundant interbeds of amphibolite 2cm to several meters thick. It also has rusty graphitic and sillimanitic schist, biotite-garnet schist, locally with abundant sillimanite, impure marble, and thin coticule beds. These rocks are sufficiently different from the Cushing east of the Norumbega Fault to warrent consideration that they may not correlate with the Cushing but may be some other, perhaps older, terrane that is a western basement for Cape Elizabeth and higher units of the Casco Bay Group.

The Cape Elizabeth Formation, metamorphosed from chlorite to K-feldspar-sillimanite grade, consists primarily of thin-bedded quartz-plagioclase-biotite (-muscovite) phyllite, schist, or gneiss (depending on metamorphic grade). Interbeds of aluminum-rich metapelite are common and, rusty phyllite and schist form mappable lenses within, and at the base of the Formation. Staurolite and sillimanite are commonly developed at proper grade. Graded bedding is infrequently observed.

The Spring Point Formation is a varied sequence of mafic metavolcanic rocks, feldspathic volcanogenic metasedimentary , and felsic metvolcanic rocks. In the western-most part of its outcrop belt at chlorite grade it is a medium gray chlorite-spessartitic garnet phyllite. Fragmental structures are not present in these rocks. At garnet grade it is a medium to dark greenish gray biotite-actinolite-plagioclase gneiss locally with distinct lineated clots of felsic material interpreted to be felsic volcanic fragments (see description for stop 5 at SMVTI, South Portland). Blue quartz phenocrysts identical to those seen in parts of Cushing are present. In the Harpswell area at the north end of Casco Bay, the Spring Point consists of dark gray garnet amphibolite at the base, and massive to thin and well bedded feldspathic metasandstone with thin interbeds of amphibolite and chloritic medium greenish gray

intermediate metavolcanics. Between the mafic and felsic rocks is a zone approximately 15m thick of garnet-rich granofels, some beds of which qualify for the name coticule.

The Diamond Island Formation is the most distinctive unit of the Casco Bay Group. It is a uniform non-bedded sequence of black, rusty-weathering graphite-quartz-muscovite phyllite characteristically with tissue-thin quartz laminae parallel to the foliation.

The Scarboro and Jewell Formations are essentially identical lithologically. They consist of rusty and non-rusty-weathering, light and dark gray phyllites of no systematic distribution within the two units. Both formations have minor medium greenish gray chlorite phyllite representing intermediate ash deposits.

The Spurwink Metalimestone separates the Scarboro and Jewell Formations. It is a 50-75m sequence of thin ribbony-bedded, medium gray fine-grained impure marble and quartz-biotite-plagioclase granofels. Similar ribbon limestone lenses less than 15m thick are present locally near the base of the Scarboro Formation, and within the Diamond Island Formation.

The Macworth Formation consists of fine-grained slightly calcareous and feldspathic medium brownish gray, commonly thinly laminated granofels with sporadic thin beds of light gray metafelsite tuff, and coarse- granule beds. The granules are white, extremely fine grained, and may be metafelsite fragments.

Regional metamorphism of the Casco Bay Group within the area of Fig.2 varies from chlorite to K-feldspar-sillimanite grade in a Buchan-type facies series (andalusite, rather than kyanite, is present in the metapelites). The present prograde assemblage is regarded to be of Acadian age (Hussey, 1985). Retrograde metamorphism, mostly recorded by the alteration of biotite and garnet to chlorite, is developed most strongly in the vicinity of faults of the Norumbega Fault Zone, and is probably genetically related to that faulting.

The Casco Bay Group is currently regarded to be of Precambrian to Ordovician age (Osberg et al., 1985) with a bias toward late Precambrian (Hussey, 1985). Brookins and Hussey (1978) report Rb/Sr ages of 481+40 Ma and 485+30 Ma for the Cushing and Cape Elizabeth Formations respectively but these probably reflect partial resetting of initial ages due to Acadian metamorphism.

The Cape Elizabeth Formation is tentatively correlated with the Rye Formation, with pelitic portions of the Massabesic Gneiss of southeastern New Hampshire, and possibly with the Ellsworth Formation of eastern Maine. The Cushing Formation, in particular the part west of the Norumbega Fault, may correlate with parts of the Massabesic Gneiss and the Cushing east of the Fault may be equivalent to the Ellsworth Formation. Units of

the Casco Bay Group above the Cape Elizabeth have no known correlations beyond their own outcrop belt.

The Cape Elizabeth Formation preserves evidence of two major and several minor deformations. The oldest folds (F1) are recumbent folds of unknown facing, vergence, extent, and age. They are known only from minor recumbent parasitic folds seen in outcrop and rare downward-facing F2 folds. F2 folds are major north-northeast-trending upright folds whose axes are gently plunging with frequent plunge reversals. F2 folds control the map distribution of the formations, except locally as at Small Point where thin, mappable units within the upper part of the Casco Bay Group have been affected by intermediate scale recumbent folding as well as later F2 refolding. The nature of multiple deformation of other units of the Casco Bay Group is essentially unknown because of the lack of well-preserved bedding.

BUCKSPORT AND CROSS RIVER FORMATIONS. The Bucksport Formation consists of thin bedded to massive medium gray quartz-plagioclase-biotite-hornblende granofels with thin interbeds of greenish gray calc-silicate granofels. Primary structures other than bedding are rare. Zones of rusty-weathering sulfidic biotite-quartz-plagioclase schist are common. Calc silicate minerals include diopside, hornblende, zoisite, and rarely grossularite. Even though metamorphosed to K-feldspar-sillimanite grade, the Bucksport Formation is not migmatized to agmatitic gneiss as are parts of the Cape Elizabeth at that grade; instead, pegmatite stringers from a few centimeters to several tens of meters cleanly cut the formation either as sills, straight dikes, or contorted lenses. The Bucksport is similar to both the Vassalboro and the Berwick Formations. The Bucksport has been mapped along strike into the outcrop belt of the Flume Ridge Formation for which Ludman (1980) suggests a Siluro-Devonian age. On the recently issued Bedrock Map of Maine, Osberg, et al. (1985) correlate the Bucksport with the Vassalboro Formation as well as the Flume Ridge and give the age as Ordovician to Devonian. If it correlates with the Berwick Formation, then the age range of the Bucksport may have to be extended back to include late Precambrian as well. The Bucksport is overlain by the Cape Elizabeth Formation, but because of inferred discordancy in ages of the two formations, the contact is interpreted to be a premetamorphic folded thrust fault.

The Cross River Formation, exposed in two belts in the Boothbay-Bristol area, consists of two members. The upper member is a medium gray quartz-plagioclase-biotite (-garnet) granofels with, in the lower parts of the member, coarse, irregularly textured amphibolite beds. The lower and most extensive member is a rusty weathering sulfidic quartz-plagioclase-biotite-sillimanite-graphite migmatite locally grading to schist. Rafts of quartz-plagioclase-biotite granofels and amphibolite probably representing non-migmatized

metasandstone beds which have been torn apart during plastic mobilization of the migmatite, are common. The contact between the Bucksport and Cross River Formations may be a premetamorphic thrust or possibly an unconformity. It is similar to rusty phases of the Penobscot Formation of Osberg and Guidotti (1974) but also resembles rusty and non rusty schists in the Cushing Formation west of the Norumbega Fault in the Gardiner-Freeport area. Resolution of this critical correlation, which has considerable implications to terrane analysis and tectoncis in southwestern Maine, awaits future field work.

INTRUSIVE HISTORY OF SOUTHWESTERN MAINE

The intrusive rocks of southwestern Maine range in age from Early Devonian to Cretaceous. The oldest rocks are plutons composed of two-mica granite, biotite granite, and granodiorite. An early Devonian age (Table 1) is reported by Gaudette, et al. (1982) for the Webhannet pluton (Figure 1), and similar ages are inferred for the many small plutons in the outcrop belt of the Casco Bay and Shapleigh Groups. The Biddeford Pluton, composed of generally non-foliated biotite granite, is of early Carboniferous age (Table 1). Carboniferous ages are also reported for the Sebago and Saco Plutons (Hayward and Gaudette, 1984; Gaudette et al., 1982; and Aleinikoff, 1984). The age for the Saco Pluton is anomalous. The diorite that makes up the bulk of the pluton is pervasively lineated (strongly) and foliated (weakly), and the original primary igneous mineralogy has been replaced almost completely by secondary minerals either by extensive deuteric alteration, or more likely recrystallization during a metamorphic episode, most likely the Acadian. Younger intrusives include alkaline ring complexes and stocks ranging in age from Permian or Triassic age to Cretaceous. Basalt, diabase, and, to a lesser extent, lamprophyric dikes are abundant in a belt about 5 km wide from the vicinity of the mouth of the Saco River southward along the coast to Kittery and beyond in New Hampshire.

In southeastern New Hampshire, Gaudette et al. (1984) report a 473 Ma Rb/Sr age for the Exeter Pluton which intrudes the Kittery and Eliot Formations post-tectonically, and after the earliest recognized metamorphism. Zartman and Naylor (1984) report a Zr age of 455 Ma for the Newburyport pluton that intrudes the Kittery Formation in southeastern New Hampshire and northeastern Massachusetts. As noted earlier, these dates appear to pose serious restraints on correlation of lithically similar belts of rock nearly on strike with each other (Berwick and Vassalboro Formations).

TABLE I. RADIOMETRIC AGES REPORTED FOR SOME PLUTONIC ROCKS IN SOUTHWESTERN MAINE

PLUTON	METHOD	AGE MA	REFERENCE
Webhannet	Rb/Sr Whole Rock	390+10	Gaudette et al.(1982)
Webhannet	Zir Pb	403+13	Gaudette et al.(1982)
Lyman	Rb/Sr Whole Rock	322+12	Gaudette et al.(1982)
Biddeford	Rb/Sr Whole Rock	344+12	Gaudette et al.(1982)
Saco	Rb/Sr Whole Rock	307+20	Gaudette et al.(1982)
Sebago	U/Pb zir	325+3	Aleinikoff (1984)
Sebago	Rb/Sr Whole Rock	325+-	Hayward and Gaudette (1984)
Exeter	Rb/Sr Whole Rock	473+37	Gaudette et al.(1984)
Newburyport	Pb-Pb Zir	455+15	Zartman and Naylor (1984)

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ITINERARY (FIGURE 2)

ASSEMBLY POINT: Commuter parking lot at exit 2 (Wells) Maine Turnpike. If there is insufficient space here, park also beside the kiosk opposite the exit to Route 109. Departure time from this lot is 8:30 A.M. We must leave promptly because of the tide at the first stop. We will not return to this point at the end of the trip.

Mileage

0.0	Turn left from exit of Maine Turnpike onto Route 109.
1.6	Junction, Route 1; turn right on Route 1.
3.6	Turn left onto Eldridge Road.
4.9	Turn left on Webhannet Drive.
5.0	STOP 1. MOODY POINT, WELLS (Figure 3).

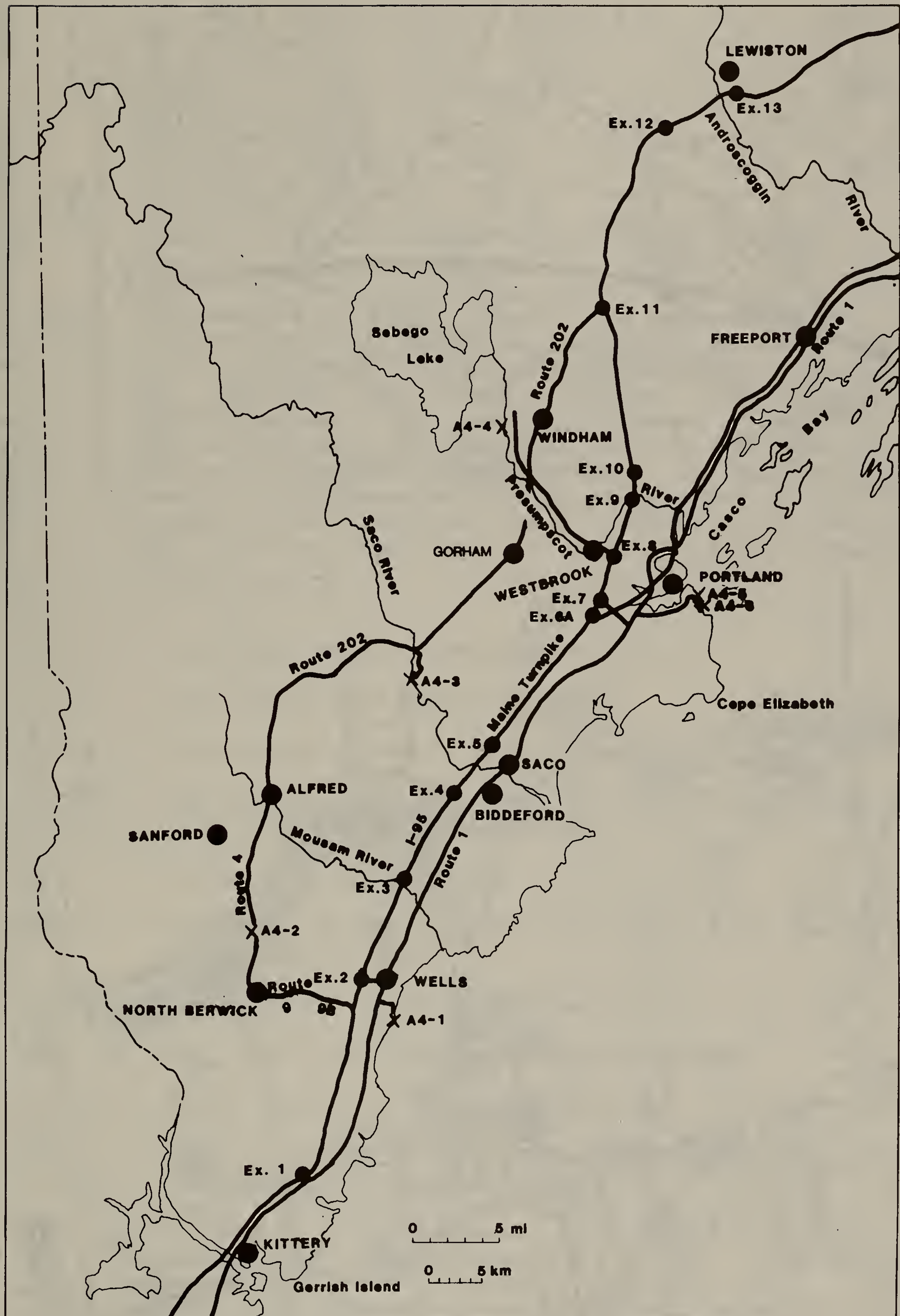


Figure 2. Itinerary and general location of stops, trip A4.

Figure 3. Location of Stop 1, Moody Point, Wells, Maine (Wells 7.5′ quadrangle).

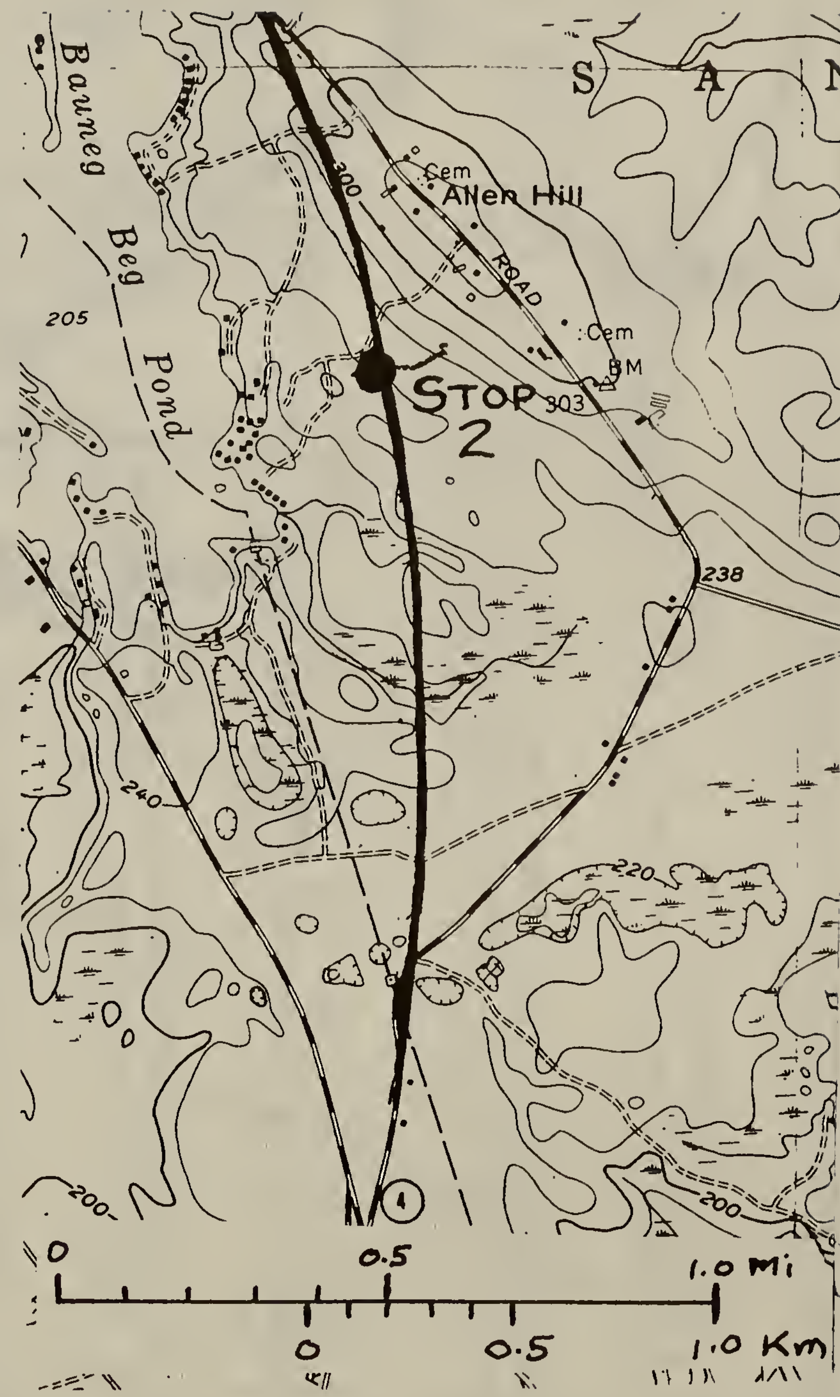


Figure 4. Location of Stop 2, Route 4 at entrance to Camp Wauban (North Berwick 7.5′ quadrangle).

Park opposite the Grey Gull Inn. We will examine exposures along the shore around Moody Point to the south. These ledges expose the Kittery Formation of the Merrimack Group and numerous diabase dikes of Mesozoic age. The Kittery Formation has been metamorphosed to only chlorite grade here. The Kittery is variably bedded, 1 cm to 2+ m. Graded bedding is common, cross beds, rare. Parallel laminations are common in some beds, absent in others. Coarse sand- to fine granule-sized grains frequently occur at the bases of thicker beds. Soles of a few beds show excellent flute casts. Four anticlinal F2 fold hinges are exposed, but the intervening synclines are covered by cobbles. These folds are upward facing, have axial planes that dip NW, and axes that plunge gently SW. The folds verge NW. Cleavage in the more pelitic beds dips gently to moderately to the northwest and is not parallel to the axial planes of the F2 folds. To what folding event this cleavage is related is not known. The psammitic parts of beds have considerable ankerite and calcite imparting to some a buff weathering color.

Continue ahead on Webhannet Drive.

5.2 Turn left onto Eldridge Road.

6.3 Turn right on US 1.

6.5 Turn left on Me 9B. CAUTION: THE 30 MPH SPEED LIMIT IS FREQUENTLY ENFORCED!!

10.7 Turn left onto Me 9.

12.2 Roadcuts of the gray biotite granite of the Webhannet Pluton.

12.8 Roadcuts of Eliot (?) Formation.

14.6 Turn right on Route 4 in North Berwick.

17.7 North Berwick - Sanford town line.

18.5 STOP 2. ROADCUT ON ROUTE 4 NEAR ENTRANCE ROAD TO CAMP WAUBAN (Figure 4).

This roadcut exposes the Berwick Formation metamorphosed to amphibolite facies. The Berwick here consists of purplish gray quartz-plagioclase-biotite-hornblende granofels with light greenish gray hornblende-diopside-grossularite granofels interbeds. Compositional layering shows moderate transposition along planes parallel to the

layering.

Continue along on Route 4.

20.7 Roadcut of granite of the Lyman Pluton.

22.4 Stoplight. Continue straight on Route 4A.

26.9 Stoplight at junction with Route 202. Straight on Route 202.

27.5 Village of Alfred, shiretown of York County. Hills on left (west) are underlain by the Alfred Complex (norite, monzodiorite, and granodiorite) of Cretaceous age.

33.9 Blinker light at cutoff leading to Route 5. Continue straight on Route 202.

35.2 Junction with Route 5. Turn right at yield sign, staying on Route 202.

35.3 Outcrop of Vassalboro Formation.

36.9 Route 5 leaves Route 202. Stay on Route 202.

40.5 Junction with Route 35. Stay on Route 202.

41.1 Junction with Route 117. Stay on Route 202.

42.8 Cross the Saco River and make an immediate right turn onto Route 117. Prior to damming, the course of the Saco River here was one of the most scenic river gorges in Maine.

43.2 Turn right onto Simpson Road.

44.1 Turn right onto gravel road opposite gravel pit.

45.0 STOP 3. SKELTON DAM, UNION FALLS ON THE SACO RIVER, BUXTON, MAINE (Figure 5).

Park in the turn around area, walk down the dirt road to the left. The other road leads to the end of the lip of the dam. At the end of the dirt road is a precipitous path down over the cliff. Watch your footing as you descend down over the cliff to the ledges below.

Ledges at the base of the dam (Figure 6) expose thin- to medium-bedded quartz-plagioclase-biotite (-hornblende) granofels of the Vassalboro formation. Ledges on the opposite side of the River at the base of the dam expose the contact between the Lyman

Figure 5. Location of Stop 3, Skelton Hydroelectric Dam, Saco River, Buxton, Maine Bar Mills 7.5′ quadrangle).

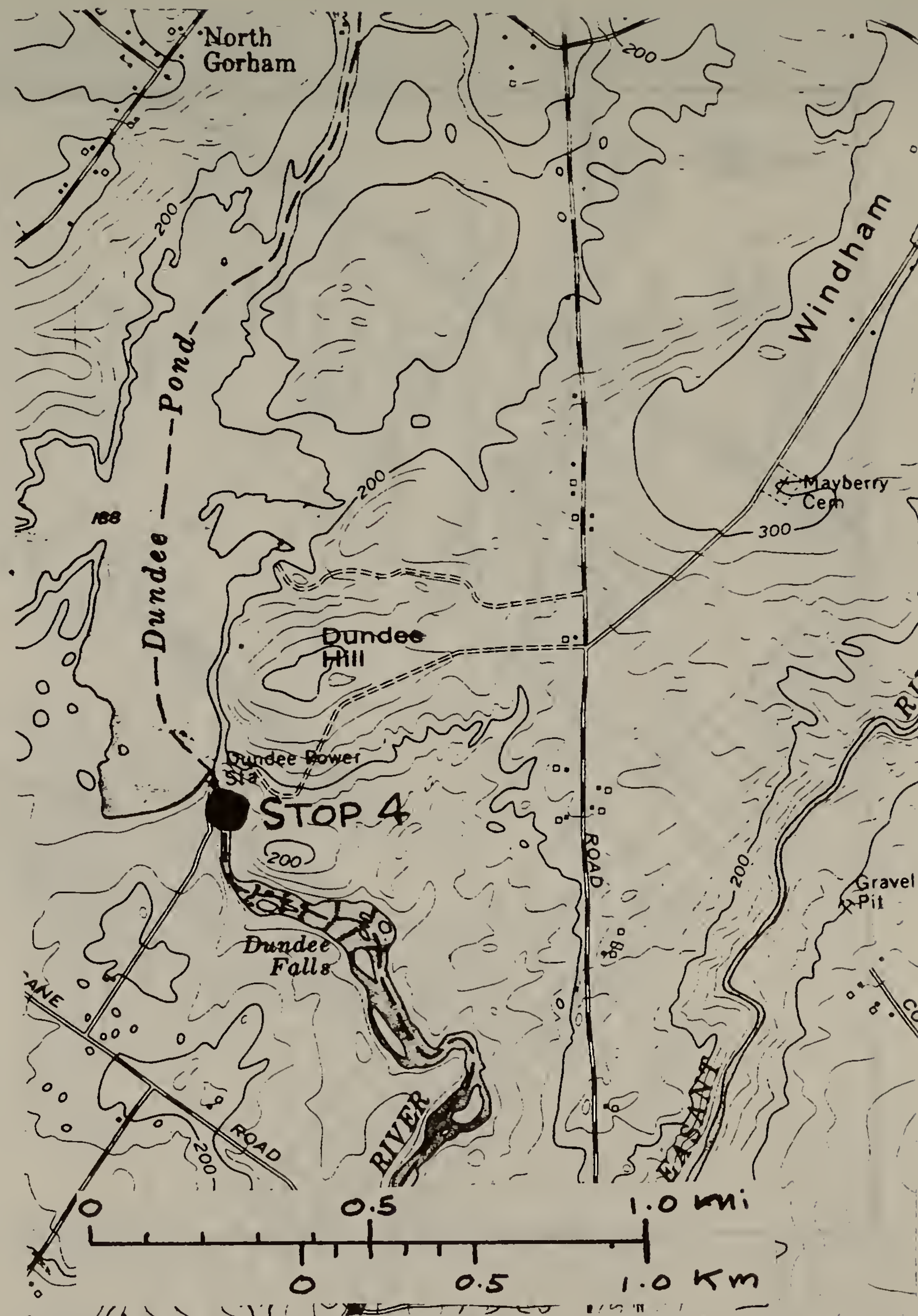
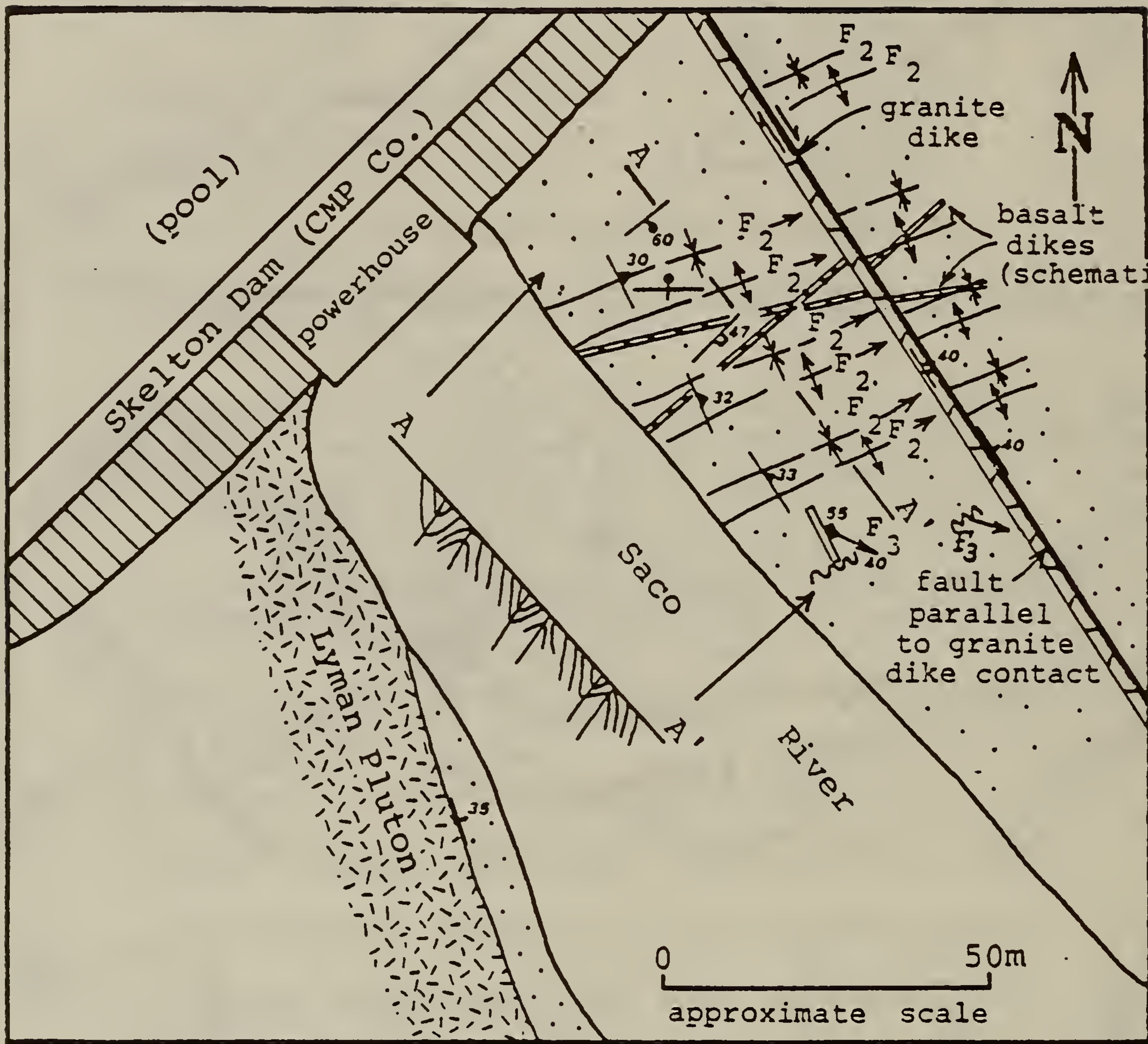


Figure 7. Location of Stop 4, Dundee Falls, North Windham, Maine (North Windham 7.5′ quadrangle.



Strike and dip of upright bedding

Strike of vertical bedding (tops in direction of ball)

Strike and dip of schistosity (S_3)

Strike and dip of granite foliation

Strike and dip of axial planes, and plunge of axes of F_3 folds

Contact showing dip where known

Strike-slip fault showing dip

Vassalboro Formation

Two-mica granite

Figure 6. Geological sketch map of the ledges exposed at the base of the Skelton Dam, Saco River, Buxton.

Pluton of Mississippian age. On the east side where we are, the Vassalboro is cut by two prominant 2-mica granite dikes probably related to the Lyman, and several basalt/diabase dikes of probable Mesozoic age.

Two fold episodes are preserved in the Vassalboro Formation. Mesoscopic scale slightly overturned parasitic folds seen in both the cliff and ledges at the base of the cliff are correlated with F2 of the regional fold sequence. No folds have been yet observed that would correlate with the early recumbent folds noted elsewhere in the Central Maine Terrane. F3 folds are minor folds whose axial planes are parallel to the contact of the Vassalboro Formation with the Lyman Pluton. Note that the biotite schistosity in the more micaceous beds is not parallel to the direction of the axial planes of F2 but of F3. The rearranged schistosity can be observed up to a kilometer east of the contact; beyond that the schistosity is parallel to F2 axial planes. The F3 folds and the biotite schistosity here are related to the intrusion of the Lyman Pluton, and are of Mississippian age. F2 folds developed during the Acadian Orogeny in Early Devonian time.

The upper contact of the eastern of the two granite dikes with the Vassalboro is a fault which post-dates the intrusion of the diabase dikes. The diabase dikes have been offset right laterally 20 cm; the component of dip slip is uncertain. Weathering and erosion of gouge along this contact has produced a deep reentrant into the cliff.

Return to cars. Turn around and return to Simpson Road via the gravel road.

45.7 Turn left on Simpson Road

46.8 Turn left on Route 117.

47.2 Turn right on US 202, following it through Gorham.

50.9 Junction with Route 22. Stay on Route 202.

55.4 Stoplight at junction with Route 4A. Stay on Route 202.

55.6 Gorham Village. Junction with Route 114. Stay on Route 202.

55.9 Stoplight. Stay on Route 202

56.0 Blinker. Bear left on Route 202.

59.3 Junction with Route 237. Stay on Route 202.

59.8 Cross Presumscot River at Gorham/Windham town line.

60.6 Blinker. Turn left on River Road.

62.4 Covered Bridge Road to the left. Stay on River Road.

63.5 Curtis Road to the right. Stay on River Road.

63.7 Park Road to the right. Park and walk down the lane to the left to ledges along the Presumpscot River at the base of the hydroelectric dam at Dundee Falls.

STOP 4. DUNDEE FALLS KYANITE LOCALITY (Figure 7).

All three units of the Windham Formation are exposed in approximately 700 feet of pavement outcrop below the dam at this locality. Near the base of the dam, the unit consists of thin-bedded two-mica + garnet + quartz + plagioclase schist. Downstream, the unit grades into a more aluminous schist with abundant staurolite, garnet, and sillimanite. Kyanite is also present but is generally associated with quartz pods. The lithology changes abruptly further downstream (Figure 8) to interbedded calc-silicate and biotite granofels (medium- to thick-bedded), and then, finally to thin bedded ribbon limestone (fine-grained gray marble layers interbedded with thinly layered micaceous quartz schist).

The rocks crop out just north of the kyanite-sillimanite isograd and therefore are interpreted to occur within the lower sillimanite zone of metamorphism. Both pre- and post-tectonic staurolite porphyroblasts have been observed in thin sections from this locality and suggest two periods of mineral growth. The first (M1) occurred during the Acadian Orogeny which metamorphosed much of the Windham-Gorham area to staurolite grade. The second (M2), a direct result of the intrusion of the Sebago Batholith at 325 Ma, caused prograde metamorphism of the M1 assemblages to sillimanite grade. M2 metamorphism recorded here is interpreted to have occurred on the underside of the Sebago Batholith at pressures of 6 to 7.5 kb (20-25 km).

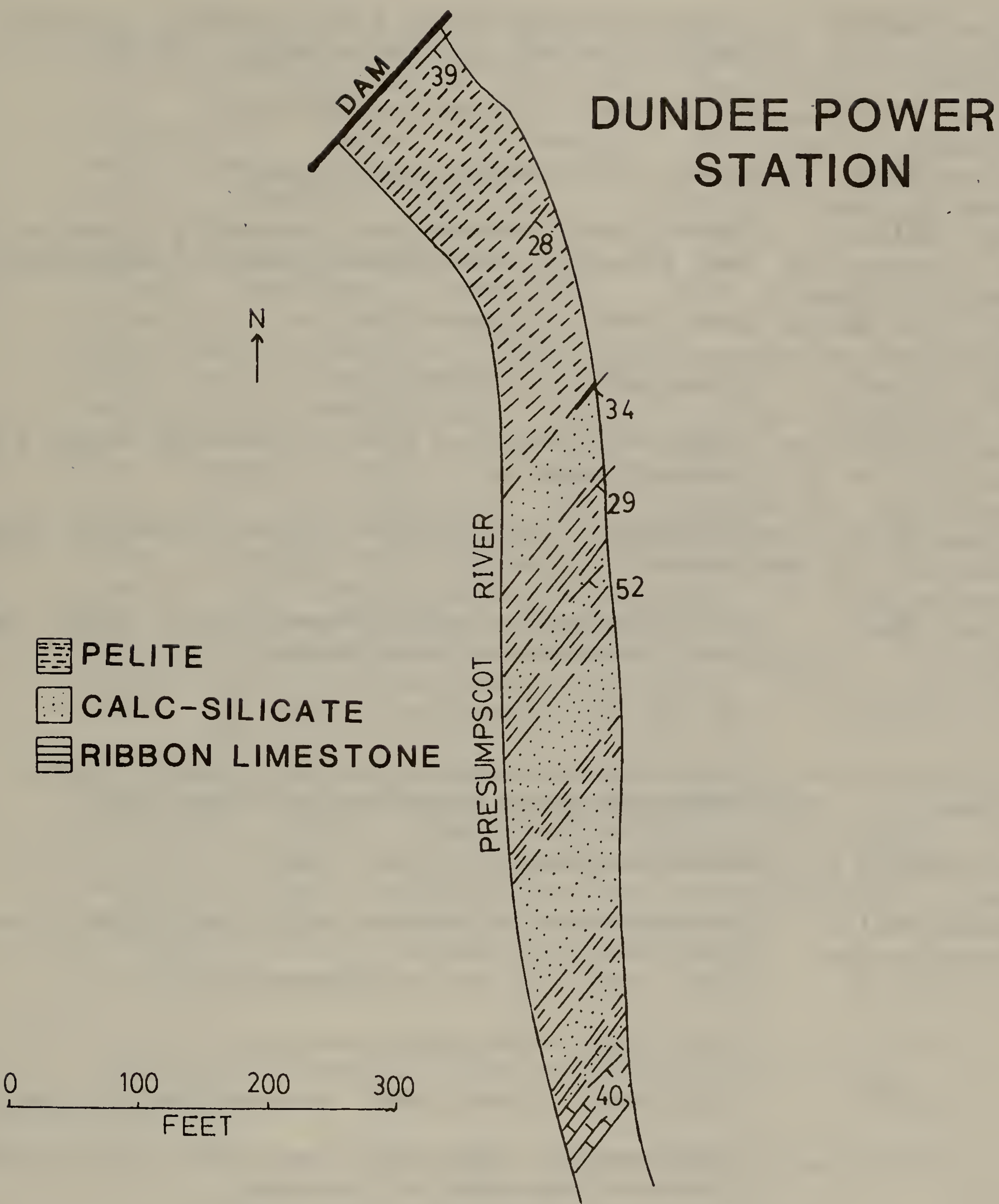


Figure 8. Stratigraphic succession at Dundee Power Station (North Windham 7.5' quadrangle), showing lithologic layers within the Windham Formation.

Return to cars, and turn around.
Return to Route 202

66.6	Junction with Route 202. Continue straight, following signs to Westbrook.
71.0	Stoplight. Continue straight.
72.5	Cross the Presumpscot River.
72.6	Turn right onto rotary (actually a series of one way streets) near paper plant in Westbrook.
72.65	Bear left in the rotary.
73.0	Stoplight at Forest St.
73.6	Turn left at stoplight, following signs for the Maine Turnpike.
74.0	Stoplight. Go straight across to Maine Turnpike booth.
74.1	Maine Turnpike booth. Immediately after, bear right for Turnpike and I-95 South.
77.9	Go off the Turnpike at Exit 7.
78.9	Toll booth (toll at writing is 20 cents). Continue straight ahead on turnpike connector.
80.4	Junction Route 1. Turn left.
81.1	Bridge over Boston and Maine Railroad tracks.
81.4	Turn right on Broadway/Route 77.
82.0	Stoplight. Continue straight.
82.2	Stoplight. Bear left, staying on Broadway.
83.6	Stoplight at Ocean Ave. Route 77 turns left. Continue straight on Broadway.
83.8	Stoplight at Cottage Road. Continue straight on Broadway.
85.0	Stopsign at junction with Pickett Street. Turn right.
85.1	Stopsign at Fort Street. Cross and enter SMVTI campus, Proceed to parking lot opposite Hildreth Hall. Walk to shoreline exposures in front of and northeast of the Hall.

STOP 5. SPRING POINT-SMVTI CAMPUS, SOUTH PORTLAND (Figure 9).

The seacliff along the east side of SMVTI campus exposes the Spring Point, Diamond Island, and Scarboro Formations of the Casco Bay Group (Figure 8). This is the type locality for the Spring Point Formation. Walk northeast to the granite embattlement (part of old Fort Prebble). The Spring Point Formation consists of medium greenish-gray actinolite-biotite-plagioclase-chlorite gneiss and schist with conspicuous 1 to 10 cm pyroclasts of felsic volcanic material. The Diamond Island Formation, one of the most distinctive units of the Casco Bay Group, is a black quartz-graphite-muscovite phyllite. Elsewhere it is frequently sulfidic and hence rusty-weathering. Close to the contact with the Scarboro is a 1 m thick ribbon metalimestone. The Diamond Island is approximately 30 m thick at this locality. This has been designated the type locality for the Diamond Island Formation (Hussey, 1985). The Scarboro Formation is exposed on the east side of the Diamond Island Formation. It consists of very weakly bedded, non-rusty muscovite-biotite-garnet-quartz phyllite. A 3 m ribbon metalimestone zone occurs very close to the base.

Ledges on the east side of the adjacent pocket beach (Willard Beach) expose the Cape Elizabeth Formation, which lies stratigraphically below the Spring Point Formation. Between the Cape Elizabeth exposures there and the Scarboro exposures here, and concealed by the sands of Willard Beach, is the South Portland Fault, a probable normal fault down-dropped to the northwest.

Return to vehicles and retrace route to the entrtance to SMVTI.

85.4 At SMVTI entrance turn left on Fort Road.

85.7 Stopsign. Bear left on Prebble Street.

86.0 Stopsign at Willard Square. Bear left, staying on Prebble.

86.6 Stopsign. Turn left on Cottage Road.

86.7 Park on Cottage Road opposite Sea View Avenue. Walk down Sea View Avenue to concrete stairs and descend to the base of the seacliff.

Figure 9. Location of Stops 5 and 6, South Portland and Cape Elizabeth, Maine (Portland East 7.5 quadrangle).

STOP 6. DANFORD COVE, SOUTH PORTLAND AND CAPE ELIZABETH (Figure 9).

Rocks of the Cushing Formation display original pyroclastic structures indicating that these rocks were originally felsic to intermediate volcanic breccia, and crystal tuff. These rocks are massive with only a minor suggestion of bedding locally. Grains of spectacular blue quartz and white plagioclase up to 3 mm in diameter are interpreted to be relict crystal fragments of an original crystal tuff. These rocks preserve weak foliation, but are strongly lineated (elongation of biotite clots parallel to the regional F2 fold hinges). Seventy five meters south of the concrete stairs the Cushing preserves distinctively recognizable breccia structure. Breccia fragments are mostly felsic and intermediate metavolcanic clasts.

Walk approximately 200 meters north from the stairs. The Cape Elizabeth-Cushing contact is exposed in the wavecut bench. The Cape Elizabeth Formation here is a biotite-chlorite-muscovite-garnet phyllite with thin interbeds of micaceous quartzite. Feeble graded bedding near the contact suggests that these beds are upright, and that the Cape Elizabeth Formation is stratigraphically above the Cushing. At this locality, the contact appears to be conformable.

This is the end of the trip. To get to Lewiston, retrace route to exit 7 of the Maine Turnpike. Travel north on the Turnpike to the Lewiston exit. Take the exit ramp for the City of Lewiston. Follow the signs for Bates College.

THE NORUMBEGA FAULT ZONE BETWEEN BATH AND FREEDOM, MAINE

Donald W. Newberg
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Lewiston, Maine

INTRODUCTION

The purpose of this trip is to examine the characteristics of several shear zones which display predominantly cataclastic textures and which collectively comprise the Norumbega Fault Zone (see Fig. 1). These structures cut two distinct lithic sequences which were earlier juxtaposed by faulting. They are the polydeformed and metamorphosed Siluro-Devonian flysch sequence of the Kearsarge-Central Maine (or Merrimack) Synclinorium and the rocks of the Coastal Lithotectonic Block (Osberg, et al., 1985) which in the area of immediate concern on this trip include the Casco Bay Group of Bodine (1965) and Hussey (1985). The contact between the two is a west dipping pre-metamorphic thrust. The location of a strong west dipping seismic reflector (Stewart, et al. 1986) coincides with that of the mapped thrust.

The trip will also examine lithologies representative of both of the above-named sequences and consider aspects of their deformational and metamorphic histories. Figure 1 is based upon field mapping conducted by the author during the last seven years with the support of the Maine Geological Survey.

Stratigraphy

The Cushing and Cape Elizabeth Formations of the Casco Bay Group outcrop primarily, but not exclusively, east of the Norumbega Fault Zone in southwestern Maine. The Cushing Formation is a predominantly metavolcanic sequence which includes quartzo-feldspathic gneisses, which are interpreted to represent felsic metavolcanics, as well as amphibolites and hornblende bearing biotite granofels which represent intermediate to mafic meta-volcanics. Metasedimentary rocks are intercalated with these lithologies. They include thin marble and calc-silicate units, rusty weathering sulfidic and graphitic schists, and metapelite. The Cushing Formation is unconformably overlain by the Cape Elizabeth Formation which includes psammitic to occasionally pelitic metasedimentary rocks. Bedding style and relict sedimentary structures, notably graded bedding, point to turbidity current deposition of the sediments of the Cape Elizabeth Formation. The persistence along strike of certain lithologies stratigraphically immediately below the Cape Elizabeth and the complete absence of basal conglomerates indicate that the unconformable relationship is a subtle - and doubtless local - one. The Cape Elizabeth and Cushing Formations appear to have shared a common deformational and metamorphic history.

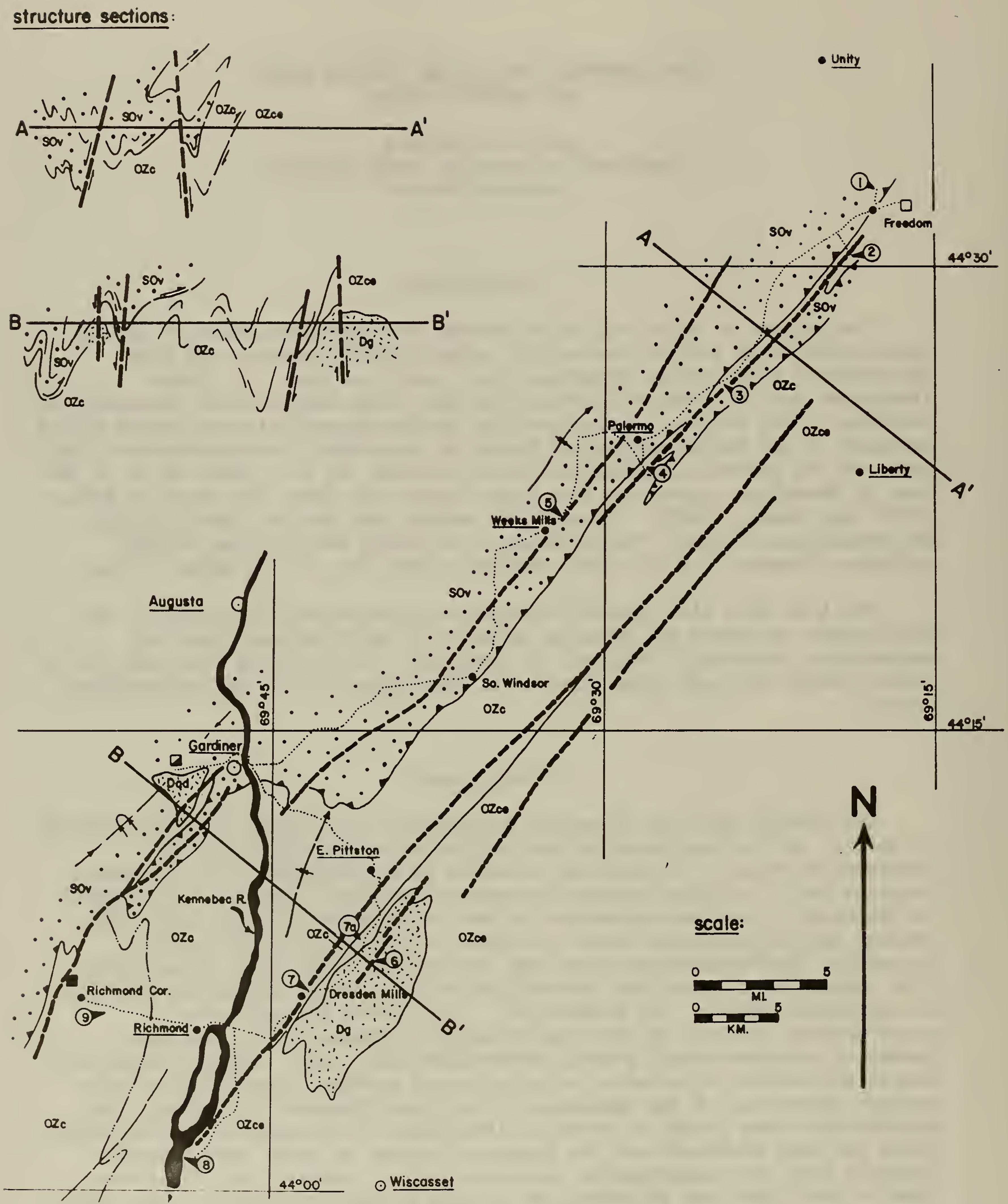


Figure 1. Geologic map of a portion of the Norumbega Fault Zone, southwestern Maine

Explanation

intrusive rocks –

Dg granite and granodiorite

Dqd quartz diorite

stratigraphic units –

SOv VASSALBORO FORMATION
calcareous metasandstone and metasiltstone (calc-silicate granofels and biotite schist)

OZce CAPE ELIZABETH FORMATION
quartz-biotite-muscovite schist and quartzite

OZc CUSHING FORMATION
felsic and mafic metavolcanics with associated volcaniclastic metasediments

symbols –

lithologic contact

trace of older fold axis

trace of younger fold axis ...with direction of plunge and dip of limbs if known

trace of pre-metamorphic fault, teeth on upper plate

trace of post-metamorphic fault

AA' structure section

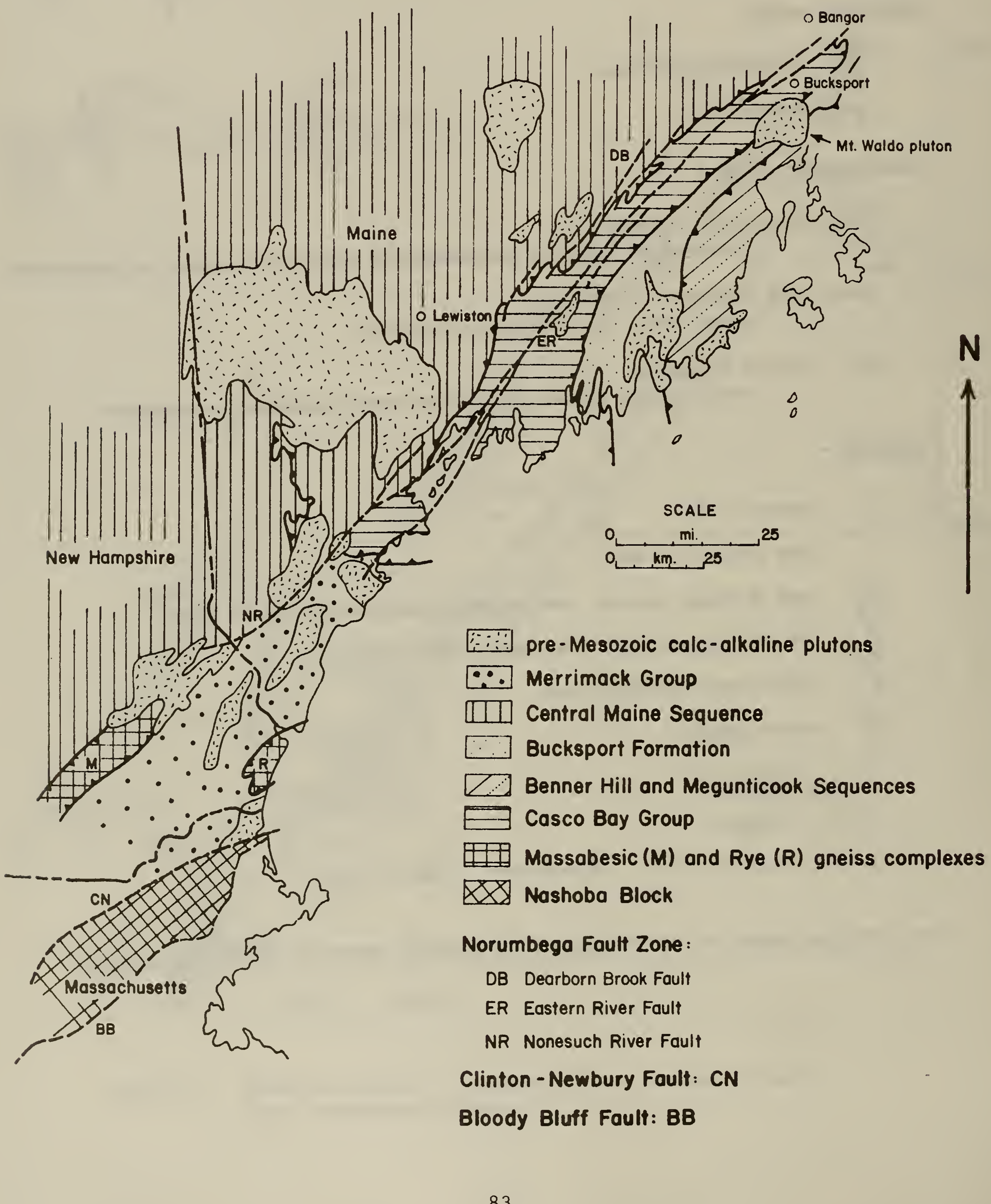
path of trip

1 trip stop

beginning of trip .... lunch stop .... end of trip

note: underlined names are also the names of U.S. Geological Survey 7.5' topographic quadrangles

Figure 2. Generalized geologic map of southwestern Maine and adjacent areas (modified from Hussey, 1985)



Faults everywhere separate the Central Maine and Casco Bay sequences. The former are, on the basis of graptolite fauna, of Silurian age and hence are considerably younger than the rocks of the Cushing and Cape Elizabeth Formations for which Rb-Sr ages of 481+/-40 m.a. and 485+/-30 m.a. have been obtained by Brookins (Brookins and Hussey, 1978).

Regional Relationships

The relationship between the Central Maine and Casco Bay sequences as discussed above is analogous in many respects to the relationship between the Merrimack Synclinorium rocks of eastern Massachusetts and the rocks of the Nashoba Block. The fault which separates them is also a west dipping structure, the Clinton-Newbury Fault (see Fig. 2). The Nashoba and Marlboro Formations of the Nashoba Block are predominantly metasedimentary and metavolcanic respectively. As such they may be lithotectonic correlatives of the Cape Elizabeth and Cushing Formations.

The lack of Acadian aged plutons and the lack of a thermal imprint on various zircon fractions from geochronologically studied Nashoba Block rocks has led others (see, for example, Zartman and Naylor, 1984; Hepburn and Munn, 1984) to conclude that the Nashoba Block represents a terrain accreted to North America after the Acadian Orogeny and perhaps as late as the Alleghenian Orogeny.

This cannot be true for that portion of the Coastal Lithotectonic Block being considered here. As shown in Figure 1 fold structures in the Cushing and Cape Elizabeth are truncated by the thrust contact and subsequently re-folded along with the thrust by folds which are northeast-southwest trending upright isoclinal folds of Acadian age. The thrust contact in Maine is therefore a pre-metamorphic fault while the Clinton-Newbury Fault is clearly a post-metamorphic structure. This difference does not negate the lithotectonic correlation noted above; but rather may illustrate the process of "diachronous welding" (Zen, 1983, p. 75).

The Nashoba Block in eastern Massachusetts is separated by the Bloody Bluff Fault from the more easterly Dedham-Milford terrane (Zen, et al, 1983) the latter having lithologic and geochronologic characteristics that suggest it is of Avalonian affinity (Hermes and Zartman, 1985). In an analogous way the region underlain by the Casco Bay Group, as well as the Passagassawakeag gneiss to the northeast, may be separated from more easterly terranes (Avalonia) by a major thrust fault in Maine (see Osberg, et al., 1985 and Fig. 2). Of interest to the question of when these Avalonian (?) terrains were accreted to North America is the fact that the fault which bounds them on the west is cut by the Mount Waldo pluton north of Penobscot Bay. The Mount Waldo has a Rb-Sr whole rock age of 390+/-10 m.a. (Brookins, 1974). A K/Ar age of 325 m.a. on biotite from the pluton was obtained by Zartman, et al. (1970). These data are consistent with the whole rock and $^{40}Ar/^{39}Ar$ mineral ages obtained by Dallmeyer and VanBreeman (1981) on three peraluminous, and presumably anatectic, granitoids which intrude the Central

Maine Sequence in the Augusta area. This consistency suggests that the postemplacement cooling histories of Acadian plutons was similar regardless of whether they intruded metasediments of Siluro-Devonian age or rocks of unknown Proterozoic Z to Ordovician age. It further suggests that the different terrains were assembled in their present configuration prior to pluton emplacement.

Norumbega Fault Zone

After the Acadian juxtaposition of rocks of the Central Maine Sequence with those of the Casco Bay Group the area was affected by the development in pre- and post-Carboniferous time of numerous shear zones displaying textures suggestive of both brittle and ductile faulting. Collectively these structures comprise the Norumbega Fault Zone. The zone bears a name first introduced by Stewart and Wones (1974, p. 230) and applied to a 3-400 m. wide zone of deformation trending N 55 E. The fault named by them in the Bucksport Quadrangle separates chlorite grade metasiltstones of the Vassalboro Formation on the northwest from sillimanite and higher grade quartz-feldspar-biotite gneisses of the Passagassawakeag gneiss (Bickel, 1971) to the southeast. The fault zone is well exposed along Route 15 between Brewer and Bucksport, where locally the texture of the Passagassawakeag has been completely destroyed. The rock is a dark ultramylonite with flinty, angular break (see Stewart and Wones, 1974, Stop #3).

Several named shear zones will be examined during this trip. Various criteria have been used to determine the sense of motion along these structures. For example, the Dearborn Brook Fault, named by Pankiwskyj (1976), follows a pronounced linear topographic low for approximately 17.5 miles. Northeast of the Village of Palermo (see Figure 1) it offsets the extrapolated Buchan type metamorphic isograds of Acadian age defined by Osberg (1971). Minor structures believed to be related to displacement along the fault indicate oblique slip that is, right-lateral, east side up, displacement. As discussed by Newberg (1985), if the orientation of the isogradic surfaces were quantitatively known displacement along the structure could be calculated. In any case displacement on the order of several hundred feet seems reasonable.

The Eastern River Fault (Newberg, 1983), whose trace passes through the Village of Dresden Mills (see Figure 1) is another linear structure within the Norumbega Fault Zone. The Cape Elizabeth Formation over an approximately 0.5 mile wide zone adjacent to the fault displays a prominent S-C mylonitic fabric. Interpretation of the microstructures in this zone indicate right lateral displacement. Offset of the Cushing/Cape Elizabeth contact as mapped in the field is consistent also with east side down displacement. The latter may be the effect of earlier ductile faulting or the effect of later brittle faulting which post-dated the emplacement of the Blinn Hill granodiorite (Newberg, 1986 and see Stop #6).

In a study of the textures associated with zones of high shear strain in several of the Acadian granitic plutons exposed east of Penobscot Bay Johnson and Wones (1984) concluded that motion was right lateral, southeast side up. This is consistent with apparent offset along the Dearborn Brook Fault but not with apparent offset along the Eastern River Fault.

Conclusions

On the basis of the relationships discussed above the following conclusions are offered:

1. The Casco Bay Group and Central Maine Sequence rocks are most likely part of the same lithospheric plate terrain ("Craton X" of Zen, 1983) but are separated by a major, pre-Acadian, west-dipping decollement.

2. The Norumbega Fault Zone includes a number of discrete shear zones having different movement histories. In general data suggest post-Acadian, pre-Carboniferous ductile faulting with right lateral strike slip displacement. This was followed by brittle faulting with dominantly dip-slip displacement.

3. From the point of view of the development of the Northern Appalachians in New England and the Canadian Maritimes cumulative displacement along the Zone is not significant and it is not a likely major terrain boundary.

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ITINERARY

ASSEMBLY POINT: Knox Corner ... the intersection of Routes 137 and 220 approximately 1.5 miles east of the Village of Freedom.

<u>Mileage</u>

0.0 intersection of Routes 137 and 220 ... proceed west on Route 137

1.0 entering Town of Freedom

1.3 bridge over Sandy Stream; cross bridge and turn right

2.1 STOP 1: outcrop of Vassalboro Formation on the east side of the road.... The rock exposed here is a strongly foliated, well sorted, calcareous metasandstone consisting of rounded 0.05-0.1 mm. grains of quartz with minor feldspar cemented by calcite. The foliation is defined by thin (>0.5 mm.) zones of biotite which show strong preferred orientation. The biotite contains abundant zircon in scattered 0.01-0.04 mm. grains. Minor detrital tourmaline and opaque minerals are present in the mica-rich intervals.

The Vassalboro here, and elsewhere in the Palermo quadrangle to the southwest shows an unusual structural feature called "asymmetric foliation boudinage" by Platt and Vissers (1980, p. 399). While ductile shortening occurred perpendicular to the foliation rigid body rotation of shear fractures and the adjacent foliation resulted in the "sausage-like" texture seen here (see Fig. 3).

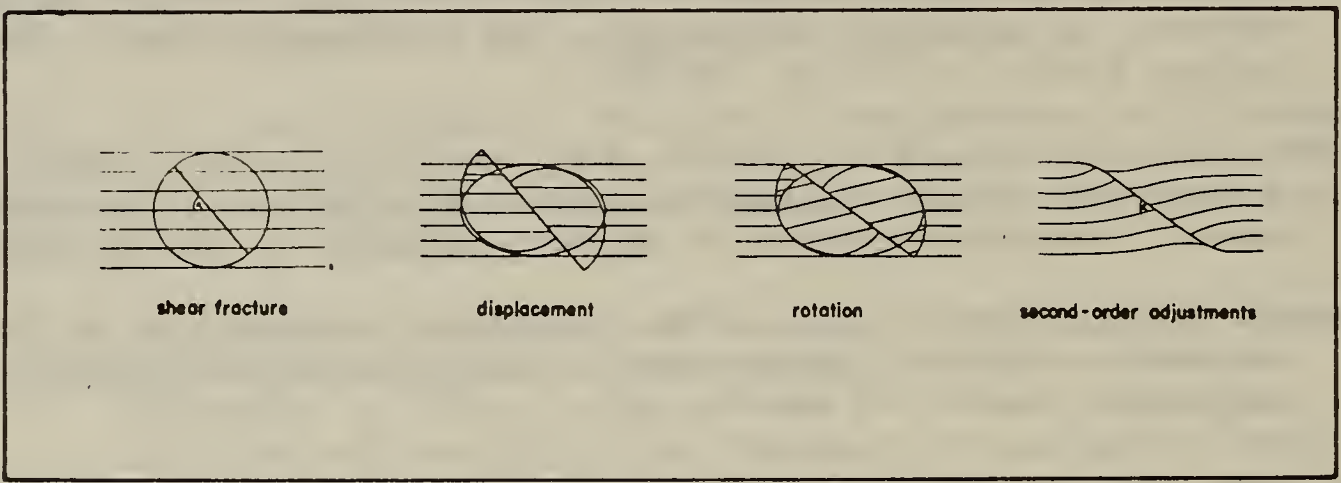

Figure 3. Progressive changes in the development of asymmetric foliation boudinage...after Platt and Vissers, 1980, p. 399.

Fluids present during deformation were responsible for the pseudomorphous replacement of biotite by chlorite + muscovite and for the precipitation of quartz in space generated by offset along the shear surfaces.

The deformation is clearly post-metamorphic and presumably related to high angle faulting along a late structure in the Norumbega Fault Zone.

turn around and return to Freedom Village

3.0 turn right (west) on Route 137

4.7 turn left onto dirt road

5.4 cross Winslow Brook.... The brook flows northeast into Sandy Pond along the inferred west dipping pre-metamorphic thrust which is the contact here between the Vassalboro Formation, exposed to the northwest, and the metasedimentary rocks of the older Cushing Formation, exposed to the southeast.

6.0 dirt road to the right, bear left

6.1 <u>STOP 2</u>: outcrops just southwest of the road along a small stream of a distinctive metapelite.... This lithology was informally referred to as the Sandy Pond Member of the Cushing Formation by Pankiwskyj (1976) who first mapped it. However, because of close spatial association with felsic metavolcanics belonging to the Nehumkeag Pond Member (informal name) of the Cushing Formation it seems preferable to include it with that member despite the distinctly different lithic character.

Thin sections of the metapelite evidence an earlier kyanite + staurolite Barrovian type metamorphism which may be pre-Acadian (C.V. Guidotti, pers. com.). The kyanite is now largely resorbed and the staurolite pseudomorphed by pinitized cordierite. The cordierite together with sillimanite, which occurs as fibrolite as well as large porphyroblasts, appear to have formed during later Buchan type metamorphism. The latter is related to the Acadian metamorphism of the Augusta-Waterville area (Osberg, 1974).

Where sillimanite porphyroblasts are abundant in the rock they stand on the weathered surfaces as resistant knots that are cut by a later cleavage. The resulting outcrop texture has lead to this lithology being dubbed "chip schist." Cordierite is pinitized particularly in proximity to post-Acadian faults.

turn around and return to Route 137

7.6 intersection with Route 137, turn left

7.6 Route 137 bears right, continue straight ahead (southwest)

11.1 Hutchins Corner (The Quebec-Western Maine Seismic Reflection line was run from Albion to Hutchins Corner and then continued to the southeast.) ...continue southwest

12.4 An approximately 10' thick limestone in the Vassalboro Formation was quarried on the left (east) side of the road presumably for agricultural purposes (?). Despite being very thin this unit is well exposed further southwest in the Palermo 7.5' quadrangle.

13.3 low outcrops of Vassalboro Formation on both sides of the road ... The same unusual boudinage seen at stop #1 may also be seen here.

14.9 STOP 3: outcrops on the east side of the Palermo (Freedom) Road of buff weathering biotite granofels (felsic metavolcanics) of the Nehumkeag Pond Member of the Cushing Formation. (This stop is the same as stop #11 of Pankiwskyj, 1978.) The thrust contact - Hackmatack Pond Fault - with the Vassalboro Formation is located approximately 300' to the west. In the field to the east scattered outcrops reflect the nature of other lithologies within the Nehumkeag Pond Member. The trace of the Palermo School Fault is located three or four hundred feet to the east. The fault is a late high angle fault with rusty weathering calc-silicate granofels of the Vassalboro Formation on its east, or downthrown, side.

17.9 outcrops of calc-silicate granofels and biotite schist of the Vassalboro Formation. These rocks are in the sillimanite zone ... in marked contrast to the very low metamorphic grade of the same lithology at stop #1 to the northeast.

18.7 intersection with old Route 3, turn left

19.6 Route 3, turn left.... The trace of the Hackmatack Pond thrust fault is inferred to pass through this intersection. To the west are units of the Vassalboro and Waterville formations of the Central Maine Sequence; to the east are exposed various lithologies of the Cushing and Cape Elizabeth formations of the Casco Bay Group.

19.9 STOP 4: The rusty weathering outcrop on the north side of the road marks the trace of the Palermo School Fault which here lies within the Cushing Formation. Thin septa of pseudotachylite are present often with chloritized and slickensided contacts. East of the fault the rock is a biotite granofels with lenses of garnet-plagioclase+/-calcite and garnet-diopside-calcite. To the west the rock is a very rusty weathering pelitic schist with pods and lenses of sheared pegmatite. Slickensides in composition surfaces oriented N60°E and dipping 60°NW rake 80°SW and indicate NW side up displacement along the structure.

turn around and proceed northwest along Route 3

21.5 Tobey's General Store

23.3 Dirigo Corner, turn left and proceed south towards Weeks Mills

23.9 outcrops of Vassalboro Formation: calcareous metasiltstone and metasandstone

26.4 STOP 5: rusty weathering sulfidic and graphitic unit within the Vassalboro Formation exposed on the east side of the road.... This outcrop is interpreted to lie along the trace of the Dearborn Brook Fault, another high angle brittle structure belonging to the Norumbega Fault Zone.

27.2 Village of Weeks Mills, turn right and proceed southwesterly

27.3 bear left

29.2 intersection with Route 32 at North Windsor, turn left (south)

34.3 intersection with Route 17 at South Windsor

note ... The trip may be continued from this point by crossing Route 17 and driving directly to STOP 6 which is located in the East Pittston 7.5' Quadrangle. The Weeks Mills, Togus Pond, and East Pittston 7.5' quadrangles would be useful, if not essential, in doing this. Hence, the log is discontinued here. Participants will travel via Routes 17, and 226 to the intersection of I-95 and Route 9-126 (just south of the Maine Turnpike) for lunch ... turn right and proceed west on Route 17.

(47.8)
0.0 parking lot on the north side of Route 9-126 opposite the truck stop and immediately west of I-95

0.7 large outcrop on the left side of the road. This is an exposure of the Mt. Ararat Member of the Cushing Formation mapped by Newberg (1984) as an inclusion (roof pendant?) within a small post-tectonic hornblende quartz diorite pluton. The exposure was also the subject of a previous NEIGC field trip (see stop #10 of Osberg, 1980)

2.1 cross Cobbosseecontee Stream, bear left and continue on Route 9-126

3.2 traffic light, turn left

3.5 traffic light, continue from right lane across bridge over the Kennebec River

3.7 traffic light at east end of bridge, turn right and continue south on Routes 27 and 126 through the Village of Randolph

5.0 turn left (east) on Route 194

5.3 small outcrop on the south side of the road of amphibolite within the Cushing Formation... This amphibolite has been traced to the southwest across the Kennebec River. It is approximately 200' thick and may represent a folded and metamorphosed mafic dike.

5.8 trace of the Dearborn Brook Fault (see Stop 5)

8.6 outcrop of the Nehumkeag Pond Member of the Cushing Formation... The rock here is a quartzofeldspathic gneiss with a discontinuous to weak foliation. This, and the other exposures of this unit are mica poor, massive and buff weathering with planar joint surfaces usually rust-stained. Often the gneiss contains scattered large garnets or irregular patches of garnet plus magnetite. The protolith is assumed to be a sequence of felsic pyroclastic rocks.

10.3 bridge over the Eastern River in East Pittston

10.5 bear left and continue on Route 194

10.9 turn right onto Nash Road

12.1 whaleback outcrop on left of amphibolite with minor calc-silicate granofels of the Cushing Formation

12.3 turn right and continue south along Blinn Hill Road... The road is located approximately 600 feet east of the unexposed contact between the Cushing and Cape Elizabeth formations.

13.6 STOP 7a: Palmer Brook, which flows west beneath the road here, exposes more or less continuous outcrop for approximately 3400'. We will walk the stream beginning by examining outcrops of garnet-muscovite-biotite schist and biotite granofels of the Cape Elizabeth Formation, located about 400' below the bridge. To the west (down stream) calc-silicate granofels of the Cushing Formation is exposed after an interval of 40' in which there is no exposure.

Amphibolite, a narrow (10-12') band of marble, and calc-silicate granofels predominate over the next half mile, which is, however, lithologically quite varied. Of particular interest are several rusty weathering sulfidic schist zones (cataclastically deformed and mineralized granofels i.e., shear zones?) and granite pegmatites, some of which show textures indication of ductile shear.

Approximately 400' west of the point where outcrop ceases in the stream there is an isolated exposure of buff weathering, finely laminated to foliated, granofels typical of the Nehumkeag Pond Member of the Cushing Formation.

continue walking west along tote road to discontinued portion of East Pittston Road (see Figure 4). From here we will return by car to the beginning of the traverse.

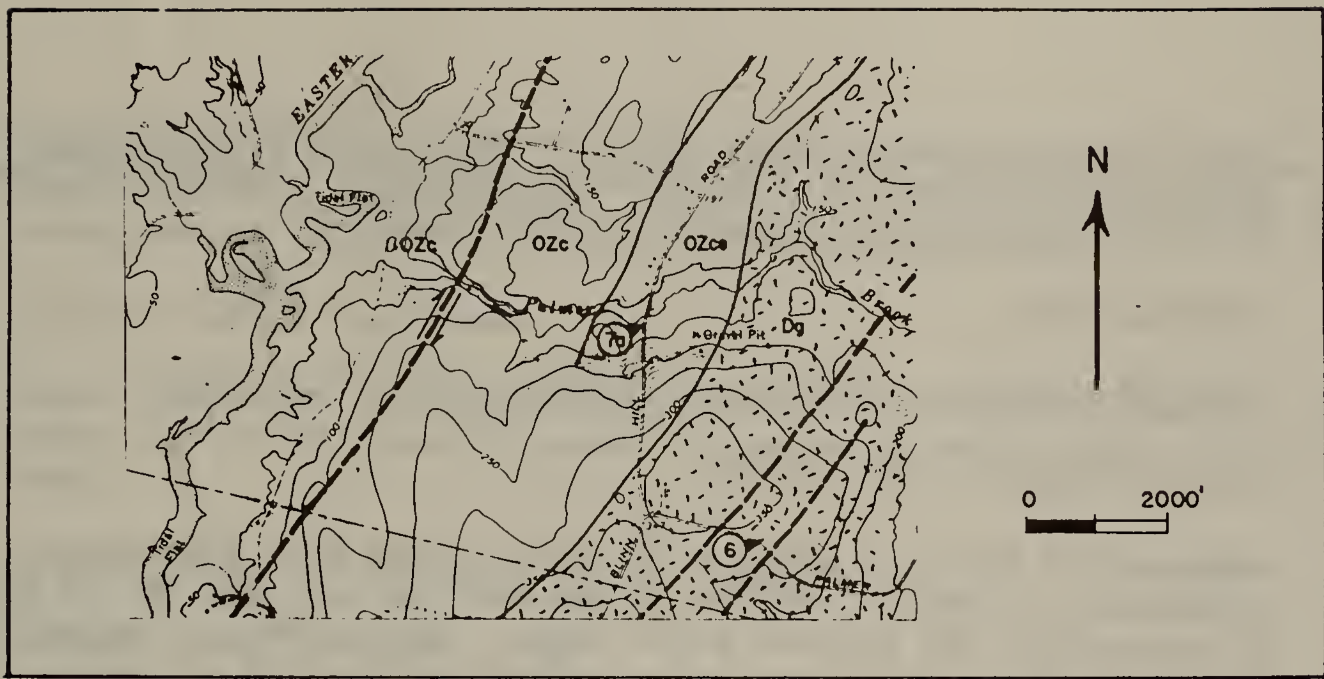

Figure 4. Location map showing stops 6 and 7a, E. Pittston 7.5' quadrangle

14.1 turn left onto Palmer Road

14.4 <u>STOP 6</u>: outcrops of Blinn Hill granodiorite on both sides of the road... A spur of the Eastern River Fault transects the pluton as shown in Figure 1. The texture seen here in outcrops north of the road is interpreted as cataclastic but relict areas of strain-free quartz seen in thin section suggest earlier ductile deformation of the granodiorite. Textures displayed by the numerous outcrops in this field are variable.

turn around and return to the Blinn Hill Road

14.8 turn left

15.1 crest of Blinn Hill, continue southwest

17.9 <u>STOP 7</u>: (The features of interest here are exposed best at low tide. Therefore the stop will be made only if the tide allows.) In the small stream flowing beneath the road and into the Eastern River just to the west there is almost continuous outcrop. Metapelite of the Cape Elizabeth Formation is in apparent conformable contact with biotite granofels of the Cushing Formation. Approximately 200' to the west a finely laminated biotite rich marble is exposed on both sides of the stream where it enters the Eastern River. These and other lithologies exposed here belong to the Wilson Cove member of the Cushing Formation (see Hussey, 1985, p. 8) As shown in Figure 1 the trace of the Eastern River Fault is located here. An indeterminate amount (perhaps as much as 24000') of apparent right

lateral displacement has occurred along the structure assuming an amphibolite exposed on both sides of the river presents the same stratigraphic horizon repeated by faulting.

continue south

18.0 intersection with Route 27 at Dresden Mills, turn right

18.1 turn left onto Route 127-197

18.7 exposure on the east side of the road of the Cape Elizabeth Formation... The 65° strike of the bedding here may be the result of drag along the Eastern River Fault. The lineation, which trends 65° and plunges 30°, is related to earlier folding or ductile faulting (?).

20.1 "Y" intersection, bear right on Route 197 and continue west

20.4 bridge over the Eastern River

21.4 exposure on the right (north) side of the road of rusty weathering granofels of the Nehumkeag Pond member of the Cushing Formation

21.6 intersection with Route 128, turn left

23.9 bridge over the Eastern River... The Cape Elizabeth Formation crops out below the bridge and to the right. The contact between the Cushing and Cape elizabeth Fromations in this area is buried beneath glacial drift, river silts, and the soils which yield Dresden's excellent potatoes.

26.7 Lincoln-Sagadahoc County line ... entering Town of Woolwich

27.7 small red house on the left, large white house on knoll on right ... turn right by a row of mailboxes onto a dirt road ... continue on road and park north of the first house ... walk west on woods road to shore at Twing Point

STOP 8: The contact between garnet-biotite-muscovite-quartz schist of the Cape Elizabeth Formation and buff weathering granofels (Nehumkeag Pond member of the Cushing Formation) is exposed here. Approximately 50' east of the contact there is a 200' thick section of amphibolite containing plagioclase, hornblende altered to chlorite with inclusions of relict pyroxene, and minor sphene, apatite, and magnetite. Narrow quartz "stripes" are seen in the outcrop in the amphibolite. In thin section these appear as cross-cutting, sub-parallel zones of strained quartz grains with mosaic texture and are interprerted as representing silica introduced along shear zones during faulting. The amphibolite is presumed to be the

same slightly discordant (?) unit that outcrops to the north at Carney Point. Approximately 100' west of the contact there is a 20' wide zone of silicified breccia within the granofels.

retrace path to main road

28.7 Route 128, turn left and continue north... There are numerous large exposures of the Cape Elizabeth Formation on both sides of the road.

29.7 County line ... entering Town of Dresden

30.2 private road to Carney Point

34.6 intersection of Route 197, turn left

35.2 bridge over the Kennebec River at Richmond

35.7 intersection of Routes 197 and 24, turn left

36.2 intersection in Richmond opposite Swan Island ... turn right on Route 197 and continue up the hill through the village

39.2 exposure on the right of the Mt. Ararat Member of the Cushing Formation. The unit is presumed to represent a sequence of metamorphosed intermediate to basic volcanics. Locally, as here, sillimanite bearing, rusty weathering pelitic volcaniclastic metasediments are interbedded with the flows (?). A late antiformal cross-fold with its axial surface dipping 30°NE and its axis oriented 320°, 20° may also be seen in this outcrop.

39.8 I-95 overpass ... park on the west side and walk south on the entrance ramp to I-95 southbound.

STOP 9: The large roadcut here is in the Mt. Ararat Member of the Cushing Formation. Texture and lithology are variable but the rocks are predominantly quartz-plagioclase-hornblende-biotite gneiss and hornblende-biotite granofels with compositional layering on a scale of 2 to 10 cms. The plagioclase averages An_{40}; occasional minor potash feldspar is present. Two samples collected a few feet from each other on the east side of I-95 at this location yielded the following modes:

apatite	2.0	1.1
hornblende	32.9	67.2
plagioclase	44.1	10.3
quartz	6.5	-
biotite	11.4	21.3
sphene	2.2	-
magnetite	0.9	0.1

There are a series of small folds here which have axial surfaces oriented 290°, 40°N. These are interpreted as being late folds. Chloritization and/or slickensiding of biotite rich composition surfaces may be due to layer parallel slip during folding. Pegmatite exposed here may be genetically related to the Sebago Pluton. The dikes can be seen cutting as well as folded by (?) these structures.

return to vehicles and continue west on Route 197

41.0 blinking light at Richmond Corner ... intersection of Routes 201 and 197

END OF TRIP

Trip B-1

FEATURES ASSOCIATED WITH THE DEGLACIATION OF THE UPPER SACO AND OSSIPEE RIVER BASINS, NORTHERN YORK AND SOUTHERN OXFORD COUNTIES, MAINE

by

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Freeport, Maine

1.0 INTRODUCTION

The purpose of this field trip is to examine some of the principal features associated with the last deglaciation in an area centered on the Saco and Ossipee River valleys in Oxford and York Counties, Maine. (See Figure 1). The area is located in the New England Upland physiographic province (Fenneman, p. 358) southeast of the White Mountains, and constitutes a physiographic boundary between the seaboard lowland and the mountains. It also marks the limit of Late Wisconsin glaciomarine sediments in the Saco basin (See Figure 2). Because of its locale, this area lends itself to the field application of some of the current working hypotheses concerning the last deglaciation. While none of the features identified in the area is unique to New England, or even to Maine, many of the associations of landforms and deposits are uncommon, and therefore invite discussions of a specific as well as a regional nature.

1.1 ACKNOWLEDGMENTS

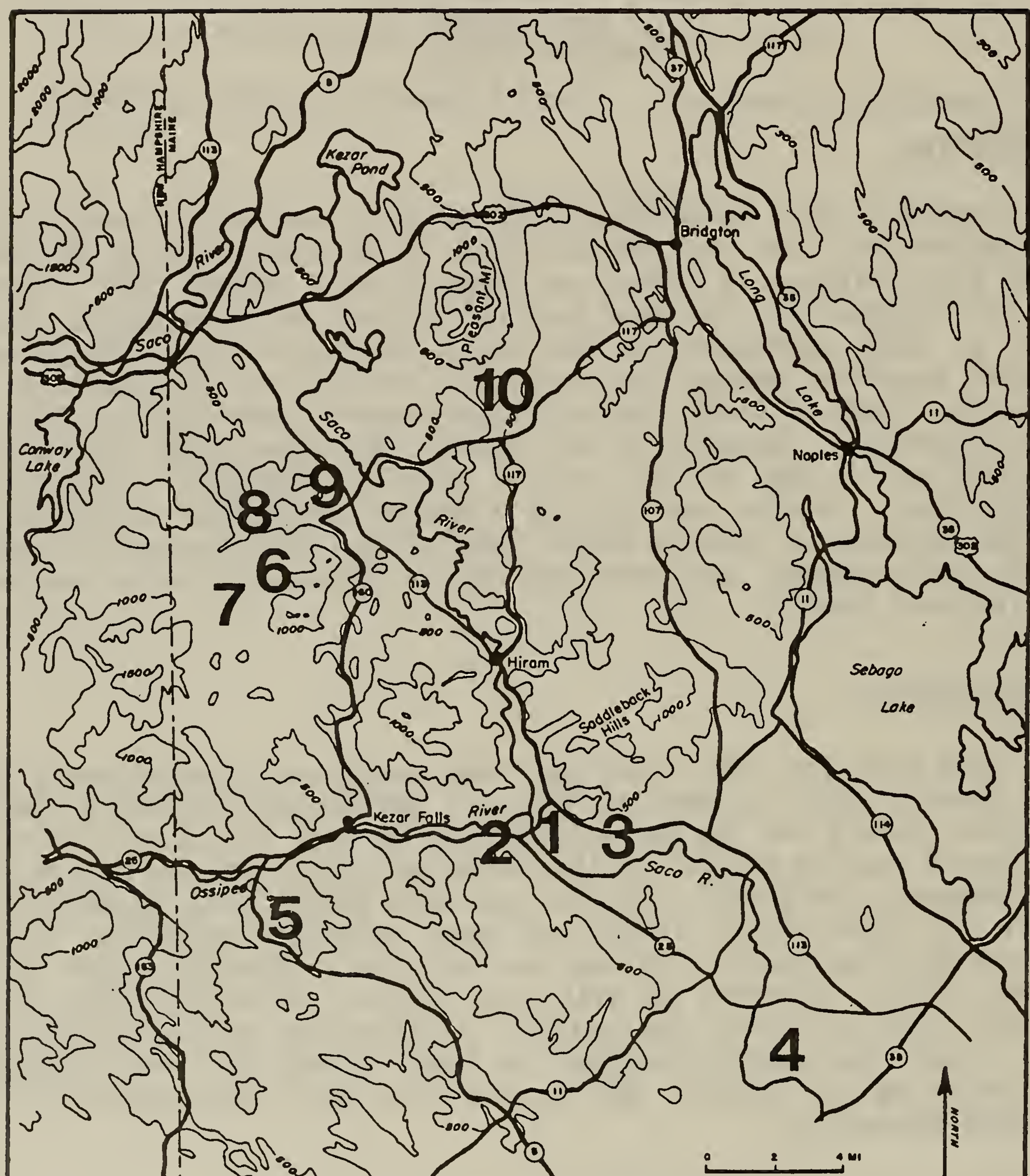
The field work for this study has been conducted intermittently during the last 4 years, both as part of the sand and gravel aquifer project for U.S. Geological Survey and the Maine Geological Survey, and as part of the on-going quadrangle mapping project of the Maine Geological Survey. My thanks go to Woody Thompson, Tom Lowell, Carolyn and Nick Eyles, Dick Goldthwait, and Carl Koteff for their field visits and many helpful suggestions over the course of the past few years. Thanks are due also to Dorothy Tepper and Andy Tolman, who kindly provided me with logs of the subsurface data produced during the recent drilling and seismic exploration of the Saco-Ossipee aquifers. I am also deeply indebted to Susan Tolman for her cartographic skills in preparing the figures, and to Jack Rand and Dick Reynolds for their review of this manuscript.

2.0 GENERAL SETTING AND TIMING OF DEGLACIATION

The southwestern part of Maine and adjacent New Hampshire are underlain chiefly by easily-weathered phaneritic plutonic rocks. In field mapping, this means that striation data are very difficult to obtain at outcrop. Because the stones derived from the local rocks are usually subrounded to rounded,

FIGURE 1

TRIP LOCATION MAP



even in diamicton exposure, fabric data are also difficult to obtain. For these reasons, little is known of the pre-latest Wisconsin history of the region. The dominant striation direction throughout the region is S15-30E, with a much fainter, younger southerly direction at times preserved. Although logic, and inference from other areas that these directions are both Late Wisconsin in age, this conclusion is equivocal. It is because of a lack of information regarding glaciation that this trip concentrates on the last deglaciation, the features of which are at least well preserved in the area. Figure 2 is a generalized map showing the distribution of the major deglaciation deposits in the region.

In the seaboard lowland of southwestern Maine, deglaciation was accompanied by a period of marine inundation. Throughout the area submerged, field evidence indicates that ice was marine-based, and that submergence and ice retreat were coeval. Ice-marginal configurations can be approximated in the area of marine submergence by following the traces of the abundant DeGeer moraines (Smith, 1982). Field evidence also indicates quite clearly that the ice was internally active while the margin stood in the marine environment (Thompson, 1982; Smith, 1982; Lepage, 1979; Thompson and Smith, 1981).

Smith (1985) established a tentative chronology of the last deglaciation of the seaboard lowland of Maine based on dates determined from seaweed, marine shells, and organic sediments collected throughout the area. The principal features of this chronology, as they relate to southwestern Maine are: 1) by 13,800 years ago, the retreating ice margin lay along the present coastline in the vicinity of Great Hill in Kennebunk, based on a date on sediments between what are interpreted to be 2 tills (Smith, 1985, p. 34). The ice margin fluctuated in a relatively restricted belt about this position for a period of roughly 600 years, as indicated by a date of 13,200 on shells in deformed marine sediments at the Kennebunk dump; 2) by about 12,500 years ago, marginal conditions lay inland of what field mapping and paleo-sea level determinations indicate was the "marine limit"; 3) by 11,500 years, the sea had regressed completely from the seaboard lowland.

This chronology is not without dispute. One of the more important questions concerns the nature of the Great Hill date. If the sediment overlying the dated fossiliferous marine silts is not in fact till, but a nonglacial diamicton, as has been suggested (B.D. Stone, 1986, personal communication), then the date would be a <u>minimum</u> date for <u>deglaciation</u>, and not a <u>maximum</u> date for the <u>presence of ice</u>. Since most of the dates acquired in the seaboard lowland are minimum dates, it is conceivable that ice marginal conditions were experienced along the present coastline much earlier than the dates would seem to indicate.

The timing of deglaciation of the area inland of the marine limit in southwestern Maine and adjacent New Hampshire is not currently established due to the lack of dates which can be tied directly to the presence of ice. Davis and Jacobson (1985, p. 358), in a regional paleoecological study, hypothesize that by 13,000 years B.P., western as well as coastal Maine, all of Vermont and all of New Hampshire were ice free, leaving a restricted ice mass in northern and central Maine.

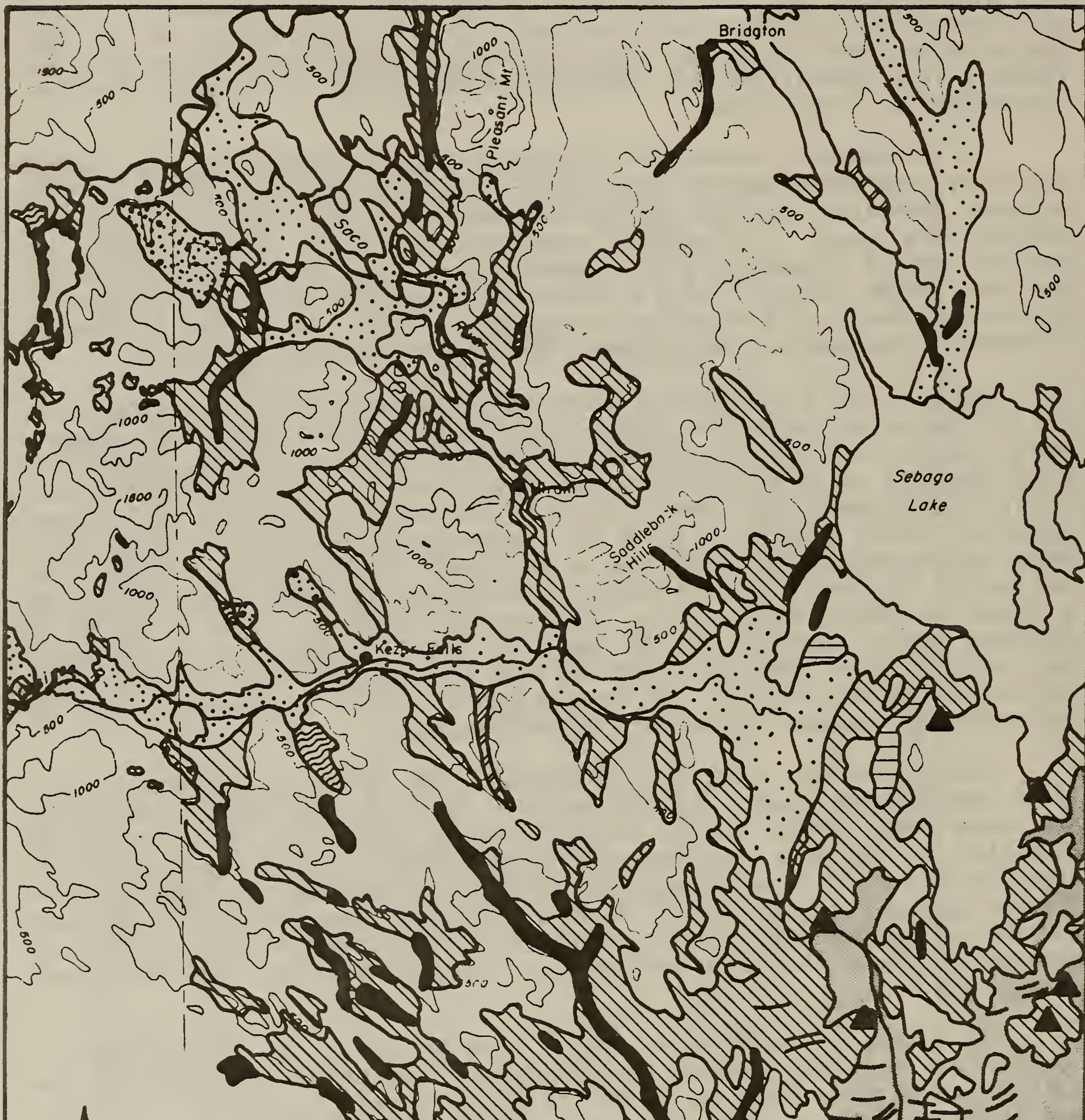

Bridgton
Pleasant Mt
Saco
Sebago
Lake
NORTH
0 2 4 MI

FIGURE 2

GENERALIZED DISTRIBUTION OF PRINCIPAL SURFICIAL FACIES

LEGEND

PROGLACIAL FACIES: Includes outwash deltas, collapsed outwash, valley train deposits, and outwash fans.

GLACIOMARINE FACIES: Includes silts, sands, and clays of subaqueous glaciomarine affinities.

SUBAQUEOUS GLACIAL LAKE FACIES: Restricted to lake bottom deposits.

PROXIMAL GLACIAL STREAM FACIES: Includes kames, kame plains, kame plateaux, kame fields, kame terraces, kame deltas, and cross-valley crevasse-fill. Does not include long-valley englacial facies.

ICE CHANNEL FACIES: Includes the principal esker systems, as distinct from ridges of sediment between kettles, and from crevasse-fill.

ICE DISINTEGRATION FACIES: Restricted to deposits composed chiefly of bouldery diamicton, with very minor volumes of stratified sediment, and having a highly irregular morphology.

SUBMARINE ICE MARGINAL FACIES: Includes the larger DeGeer moraines and subaqueous fans.

TERRESTRIAL MORAINE FACIES: Applies to so-called "ribbed moraine" of uncertain genesis.

UNDIFFERENTIATED DIAMICTON FACIES: Includes deposits of till, and all other diamicton facies not included elsewhere.

SPECIAL FEATURES

DEGEER MORAINES: Locations and numbers stylized.

EXPOSURES OF TOPSET-FORESET CONTACTS IN GLACIOMARINE DELTAS

Sources: Smith (1977a, 1977b), Thompson (in press, 1976a, 1976b), Thompson and Borns (1985), Newton (1974), and Holland, (unpub. mapping).

2.1 THE STYLE OF DEGLACIATION IN SOUTHWESTERN MAINE

"...I have long since learned that glacial rivers bear careful watching. Their deceitfulness is well exhibited..."(Stone, 1899, p.252). The style of deglaciation of southwestern Maine and adjacent New Hampshire landward of the marine limit raises questions about which there little consensus exists. Essentially similar field data have been invoked at different times and by different people as evidence for either areal stagnation, or for progressive, active retreat. One wonders whether at least part of the controversy could be avoided if everyone first agreed upon a meaningful definition of the terms "active" and "stagnant" ice for field use in interpreting ancient deglaciation environments. Such a definition would include a set of recognizable field criteria which could be linked empirically to certain ranges of ice flow velocity through observation of existing glaciers in dynamically equivalent settings. To my knowledge, the only attempt at such a definition as applied to northern New England is that proposed by Goldthwait and Mickelson (1982), using an analogy between the White Mountains of New Hampshire and the Glacier Bay area of Alaska. Using their example, I have assumed a flow velocity of 10 meters/yr to distinguish between "stagnant" and "active" ice. I have also assumed that "dead" or "disintegrated" ice implies velocities approaching or equal to 0 - like an ice cube resting on a table-top.

By way of introduction, it is perhaps more appropriate to present the question of deglaciation style of inland southwestern Maine and adjacent New Hampshire in terms of whether ice retreat involved principally vertical "retreat" (top to bottom), or horizontal retreat (front to back). Stated this way the problem becomes more workable, and there are several lines of background information which are applicable and worthy of consideration: 1) radiometric dates; 2) field geology in the highland areas; 3) field geology from the seaboard lowland; 4) field geology along the larger river valleys above the marine limit; and 5) theoretical modeling:

1) There are numerous dates from high elevations in the mountains of New Hampshire and Maine (Spear, 1981; Davis and Jacobson, 1985; Borns and Calkin, 1977). The dates closest to the trip area are from Deer Lake Bog and Lake of the Clouds (Spear, 1981), in the White Mountains of New Hampshire. The dates for these localities indicate that the high peaks of the Presidential Range were ice free by between 13,000 and 14,000 years B.P.

If the date of 13,800 from Great Hill in Kennebunk is accepted as indicating ice along the coast at that time, and if the dates from the high ponds in the White Mountains are accepted, the concept of a "thinning pancake" (Lowell, 1983) is compelled. Even without the Great Hill date, the Kennebunk dump date still puts ice along the present coastline at a time when at least part of the White Mountains has been deglaciated.

2) Recent work by Gerath, Fowler, and Haselton (1985), Gerath and Fowler (1982), Davis (1986), and Goldthwait and Mickelson (1982) has reaffirmed the interpretation by Goldthwait (1970) that ice thinned over the White Mountains fairly early during deglaciation, and that no reactivation of the White Mountain cirques occurred during "post-Laurentide" time. A similar conclusion was reached by Borns and Calkin (1977) for the Boundary Mountains of Maine, and by Davis (1983) for Mt. Katahdin.

3) There is little question from field evidence that the retreat of ice through the seaboard lowland of Maine was progressive, and involved ice which retained sufficient internal activity to produce an abundant sediment supply, to construct end moraines, and to deform sediments (Smith, 1982, 1985; Thompson, 1982).

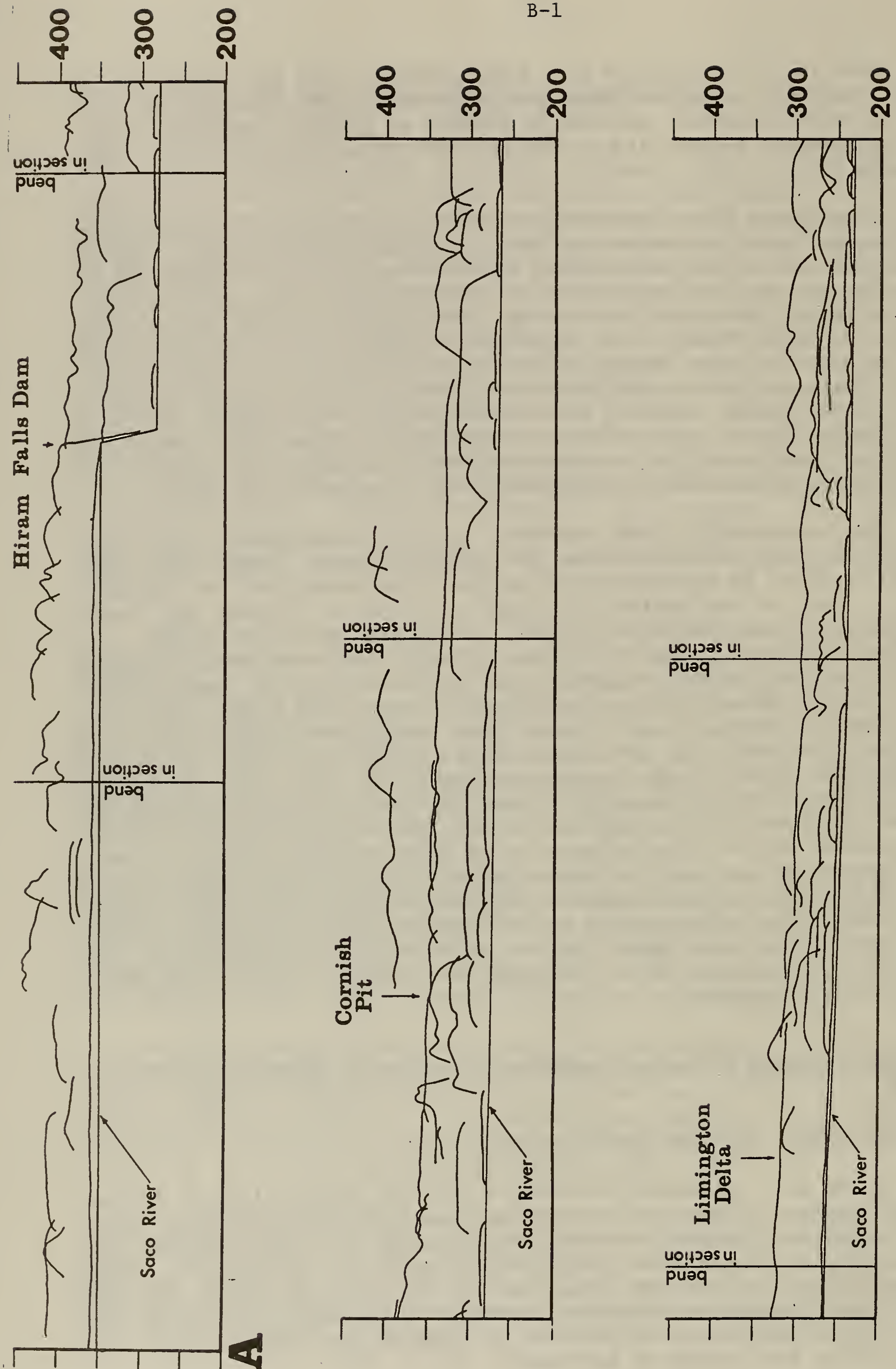
4) The evidence from above the marine limit along most of the principal river valleys in Maine is somewhat equivocal. Caldwell, Hanson, and Thompson (1985) felt that the volumes of sediment concentrated at and inland of the the Late Wisconsin marine limit argued for the presence of active ice after terrestrial conditions occurred. Similarly, the presence of the Androscoggin moraine near Gilead, Maine, and associated easterly striations and roche moutonnee in the Milan, New Hampshire area pose a convincing picture for active ice in that area during deglaciation (Thompson, this volume, Trip C-1). However, Holland (1982), working in the Machias and Narraguagus basins in Washington County, Maine, felt that after undergoing a transition from marine to terrestrial conditions, the ice mass immediately inland of the marine limit in southeastern Maine essentially downwasted and disintegrated in situ.

5) The increasingly clear evidence for the development of a late-glacial ice cap over northern Maine and adjacent Quebec (Lowell and Kite, 1986a and b) requires an explanation in terms of the dynamics of the Late Wisconsin ice. Such an explanation has been provided by Fastook and Hughes (1982), Mayewski, Denton and Hughes (1981), and Hughes and others (1985). The resulting model has been summarized by Borns (1985), and incorporates downdraw along ice streams into calving embayments on two fronts (the St. Lawrence valley and the Gulf of Maine) during the last glaciation. The model states that as the ice sheet thinned, it was cleaved along topographic highs quite early during deglaciation, with the development of a residual ice mass. Because of the high ablation rates along the marine margins of the ice, the surface profile would have been considerably flatter than the Nye equilibrium profile. Implicit in this hypothesis is the notion that following retreat of the ice from the seaboard lowland, the deglaciation of Maine was characterized chiefly by vertical thinning, and that the deglaciation of large inland areas of Maine was essentially through areal stagnation, followed by disintegration. According to the model, the logical timing of the "pancake stage" would be when the ice became permanently land-based; the ice mass would have insufficient surface gradients to support active flow once terrestrial, rather than marine ablation processes dominated.

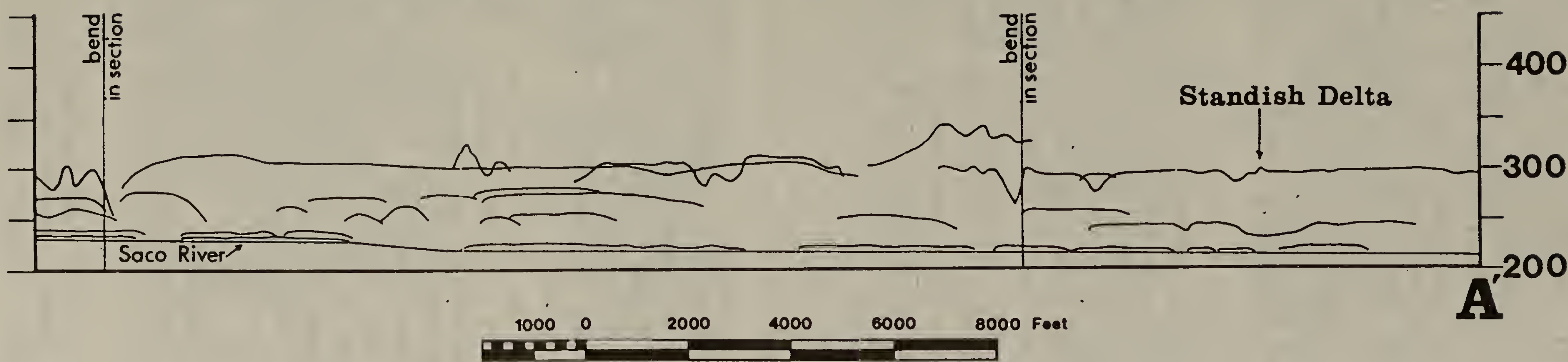
3.0 PRINCIPAL FEATURES OF THE DEGLACIATION OF THE UPPER SACO AND OSSIPEE VALLEYS

3.1 THE RELATIONSHIP WITH THE "MARINE LIMIT"

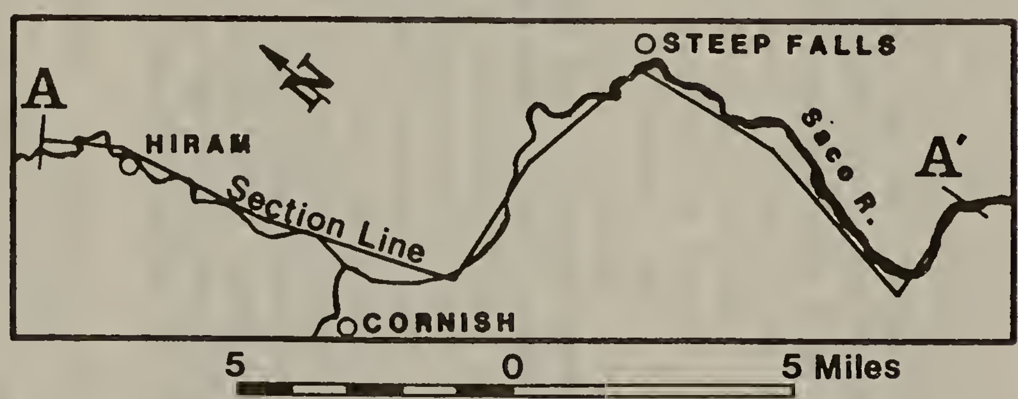
Some of the most noteworthy features of the last deglaciation in the upper Saco and Ossipee valleys are the extremely well-developed terraces lining the valley corridors. Mapping indicates the presence of several sets of terraces along the Saco River from Standish to Hiram Falls (Figure 3). These terraces range in elevation from less than 10 feet, to well over 100 feet above the present floodplain. Terraces separated by as much as 90 to 100 feet of relief, from the highest kame terraces to some of the lowest fluvial terraces, all display some degree of kettling.



A
400
300
200
Hiram Falls Dam
bend in section
bend in section
Saco River
400
300
200
Cornish Pit
bend in section
Saco River
400
300
200
Limington Delta
bend in section
bend in section
Saco River



SECTION LOCATION MAP



NOTES:

1) Vertical scale on sections is in feet above NGVD.
2) Profiles constructed by orthogonal projection of the highest surfaces of stratified deposits on either side of the section line.
3) Sections read from north to south, from the top to bottom in the figure.

FIGURE 3

TOPOGRAPHIC PROFILES OF STRATIFIED DEPOSITS IN THE SACO RIVER VALLEY

STOP 1 is an exposure in the most prominent terrace set in the Cornish area. The feature is part of a relatively continuous set of terraces which grade into an apparently glaciomarine delta in Limington (Figures 2 and 3), over which we will pass en route to **STOP 3**. The fluvial bedforms displayed in this pit are typical of the terraces correlated with the delta along the Ossipee River into New Hampshire, and along the Saco River as far as Hiram Falls. The deposits as a whole are considered to be collapsed outwash, resulting from the deposition, from sources far upstream, of sand and gravel over a thin, discontinuous assemblage of relict, disintegrating ice masses lying in the Saco and Ossipee valleys. A water plane associated with these terraces may be easily drawn from the delta up to Hiram Falls, at which point the surface elevations of stratified drift increases rather abruptly (Figure 3).

The delta at Limington is the furthest inland of all the glaciomarine deltas in the Saco basin. The morphological relationship between the distal part of the Limington delta (an outwash delta), and the proximal part of the next delta downstream (the Standish delta, which is an ice-contact delta; **STOP 4**) indicates that the Limington delta was formed last, and suggests that it therefore marks the maximum marine limit in the Saco basin. If a zone of relatively thick, continuous glacial ice existed in the area at the time of the maximum marine submergence, it was upvalley from the delta some distance- at least to Hiram Falls, and possibly considerably further north. This is indicated by the fluvial and proglacial nature of the terrace deposits and the association with the Limington delta.

3.2 THE DEVELOPMENT OF GLACIAL LAKES

A separate issue, but one related intimately to the interpretation of the terraces at Cornish, involves the initial development of the proglacial lake(s) which occupied the Saco and Ossipee valleys during the last deglaciation. Conformably underlying the deposits which constitute the fluvial terraces along the Saco and Ossipee rivers are clays and silts. These sediments, which are known principally from engineering boring logs, residential well logs and U.S.G.S. aquifer exploration boring logs (Prescott, 1979; D.Tepper, personal communication) are generally continuous up the Saco valley from Hiram Falls to Bartlett, New Hampshire, and up the Ossipee valley,at least to the Ossipee Lake basin (Newton, 1974). The sediments are generally quite deeply buried by sands and gravels, and outcrop at the surface in only a few places. Leavitt and Perkins (1935) mention that exposures of "varved" fresh water clays beneath the sand plains of Fryeburg and Conway were seen during their work in the area, but no descriptions of the locations were included in their report. Neither W.B. Thompson (personal communication), who is currently mapping in the Fryeburg area, nor I have yet rediscovered these exposures. By far the best surface exposure of seemingly similar sediments of which I am aware is in the village of Cornish (**STOP 2**). At this locality, the deposits are rhythmically bedded, and appear to be lacustrine. The deposits outcrop at approximately 325’, which is 40’ to 50’ below the elevation of the surface of the superadjacent fluvial terrace.

An important question is whether the lake-bottom sediments in the upper Saco valley were deposited into a single large, long-lived glacial lake, or whether there was a series of smaller glacial "ponds", each with a separate

base-level control and a unique water plane. This question is difficult to answer directly; the Saco is a south-draining river, meaning that the only possible dams in the Saco valley would be either relict ice and/or drift, and as of this writing, I have found no irrefutable dams or spillways. However, in certain areas of the Saco-Ossipee corridor, such as near the village of Cornish, a potential answer to the lake problem may be deduced.

The stratigraphy of the deposits at Cornish indicates that in the area of confluence between the Saco and Ossipee Rivers, the lacustrine episode preceded the last glaciomarine episode. The fluvial terraces grading to the Limington delta are consistently kettled (indeed, even lower terraces are often kettled), indicating that the valley was far from ice-free prior to terrace formation. This means that the clays exposed at Cornish were deposited in the initial "stage" of what was ultimately a somewhat larger glacial lake. This initial stage featured a very narrow east-west strip of open water, which was impounded by a drift- or ice dam across the valley 0.2 miles upriver from the Limington delta (**STOP 3**). Because of the lack of deltas, spillways or shoreline features associated with this stage, its water-surface elevation is unknown. It must be higher than 325' (the elevation of the bottom sediments at Cornish), but no upper limit can be found. Presumably, the lack of water-plane indicators suggests that the initial stage of the lake was short-lived.

There are two alternative hypotheses that can also accommodate the field data:

1) The rhythmically-bedded deposits beneath the terrace in Cornish are not lacustrine, but marine (or estuarine). This obviates the need to hunt for a dam and spillway, but creates a separate problem: at what point, and on what genetic basis, would a boundary be established with the clay/silts (which are certainly lacustrine) further up the Saco valley ?

2) The Limington delta is not glaciomarine, but a glaciolacustrine delta, built into a lake dammed by ice or drift further downvalley; perhaps by the Standish delta, which is definitely glaciomarine. A precise determination of the water plane associated with the Limington delta is required in order to determine whether hypothesis is appropriate.

3.3 MORAINES

Of the several pieces of evidence which are conventionally used to document the presence of internally active ice during deglaciation, one of the more sought after is a moraine. If it can be demonstrated, as has been done in the seaboard lowland, that a particular morainic feature or assemblage is an ice-marginal phenomenon, and if there is evidence of ice-shove deformation within the feature, then the picture is clear. On the other hand, the mere presence of a morainic landform (as a morphologically distinct and independent entity) does not by itself imply active ice, nor does it necessarily imply an ice marginal environment. Several types of morainic landforms in the Scandinavian countries have demonstrably subglacial origins (Lundqvist, 1969). Two uniquely different types of morainic topography exist in the upper Saco-Ossipee basin.

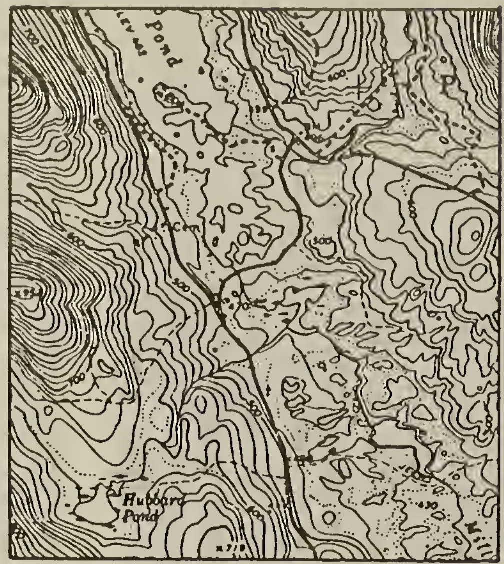


FIGURE 4a

FIGURE 4b

From U.S. Geological Survey Kezar Falls 7.5' Topographic Quadrangle Map

CONTOUR INTERVAL 20 FEET

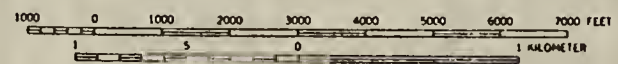
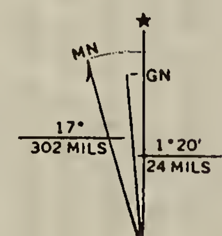
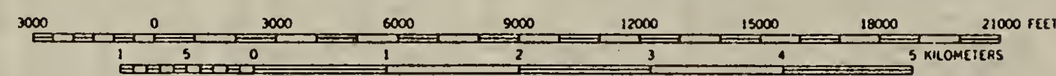


FIGURE 4c



From U.S. Geological Survey Ossipee Lake 15' Topographic Quadrangle Map



CONTOUR INTERVAL 20 FEET
DATUM IS MEAN SEA LEVEL

FIGURE 4

MORPHOLOGY OF "RIBBED MORAINE"

3.3.1 "RIBBED" MORAINE

The first type includes independent linear ridges, composed exclusively of coarse-grained glacial diamicton. Within the study area, the ridges are found exclusively in the bottoms of second- or third- order tributary valleys whose trends range from N0E-S0W to N30W-S30E, subparallel to what are most likely the 2 most recent directions of glacier flow, as inferred from striations. The length:width ratios of these moraines generally exceed 4:1, but are less than 10:1, making the moraines noticeably "stubbier" than the coastal DeGeer moraines. The long axes are oriented generally normal to the long axes of the containing valleys; the moraines occur exclusively in clusters, or "swarms"; and they have been found in both north-draining and south-draining valleys.

Whether the repetitive morphology of the moraines is an indication that internally active ice was involved in their formation is open to question. Even if active ice was responsible, however, there is still the question of the environment of deposition of the moraines. Preliminary field work indicates that in those cases where proglacial stratified sediment coexists with the moraines in a given valley, it overlaps them on both their distal and proximal faces, suggesting that the moraines might not be marginal features, but submarginal or subglacial. Although it is possible that the overlying proglacial sediments may have been produced sequentially upon retreat of ice to successive marginal positions (each of which would be marked by a moraine "line"), the uniform texture of the stratified deposits from the proximal and distal sides of a single moraine would seem to argue against this interpretation. If the moraines are subglacial, then whether or not they were formed by active ice becomes a moot point when considering them as indicators of deglaciation style; they could have formed very early during deglaciation, when the ice was still relatively thick.

Figure 4 shows the appearance of several of these moraine localities on both the 15-minute and 7.5-minute quadrangle map scales. Superficially similar moraines have been identified elsewhere in Maine (Lowell, 1981; Caldwell, 1976; Holland, 1983; Thompson and Borns, 1985), but a consistent and uniform interpretation of their genesis has not yet appeared. As Caldwell, Hanson, and Thompson (1985) observe, the regional distribution of the features appears to be restricted, as it is in this area, to terranes of phaneritic plutonic rock. The features also bear at least an initial resemblance to the so-called "Storso moraine" described by Lundqvist (1981) in Scandinavia. Because of the current lack of understanding concerning the origins of these features, the reminology presently used to map them in Maine is simply "ribbed moraine".

STOP 5 is in one of the most accessible and best exposed of the ribbed moraine localities in the region. It is hoped that the evidence presented at this stop will generate discussion, because the interpretation of these moraines is one of the keys to forming a working hypothesis concerning the mode of deglaciation of the region.

3.3.2 "ICE-DISINTEGRATION MORAINE"

The second type of morainic deposits exhibit an extremely hummocky, essentially random topography, and is generally found on valley sides above

FIGURE 5a

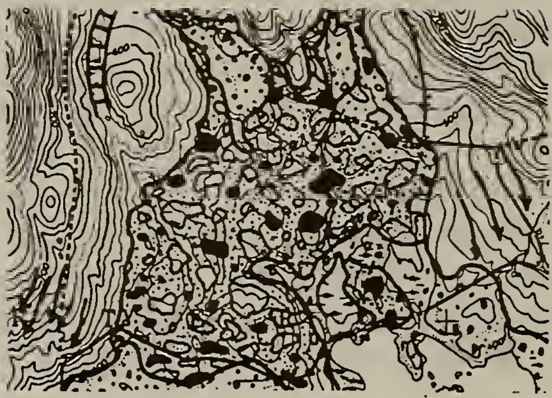


FIGURE 5b

Topography From U.S. Geological Survey Kezar Falls 7.5' Quadrangle Map

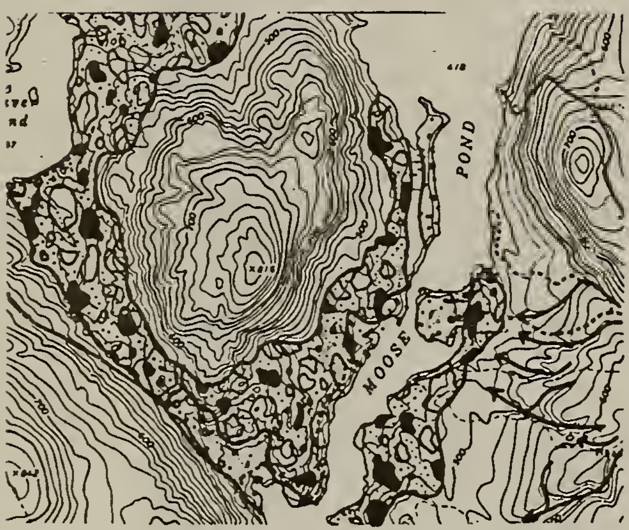
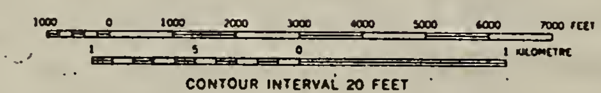


FIGURE 5c
Topography From U.S. Geological Survey
Hiram 7.5' Quadrangle Map

FIGURE 5d
Topography From U.S. Geological Survey
Cornish 7.5' Quadrangle Map



LEGEND

BOULDERY DIAMICTON FACIES

PROXIMAL GLACIAL STREAM FACIES

DISTAL GLACIAL STREAM FACIES

MELTWATER CHANNELS

FIGURE 5

ICE DISINTEGRATION FEATURES

the locally highest stratified deposits. The deposits are composed largely of sandy, bouldery diamicton, but also contain minor intervening "pockets" of stratified sediment, exposed both at the surface and in cuts. There is almost always a close spatial relationship between the deposits and lateral meltwater channels (see Figure 5). In many instances, there is a deposit of continuous stratified sediment which extends downvalley away from the morainic debris. Most of the stratified deposits exhibit some degree of collapse. The occurrences of stratified sediment in association with the morainic debris appear to be restricted to south-draining tributary valleys.

I have classified the deposits tentatively as "ice-disintegration moraine" because of the texture of the sediment; the position of the deposits relative to the principal deposits of stratified drift in a given valley (as contrasted to the relatively minor volumes of stratified drift included within the morainic mass); and the close association with well-developed meltwater channels, and the similarity to features described by others elsewhere (Boulton, 1967; Gray and Lowe, 1977; and Sissons and Sutherland, 1976) as representing superglacial deposition in an essentially dead-ice environment.

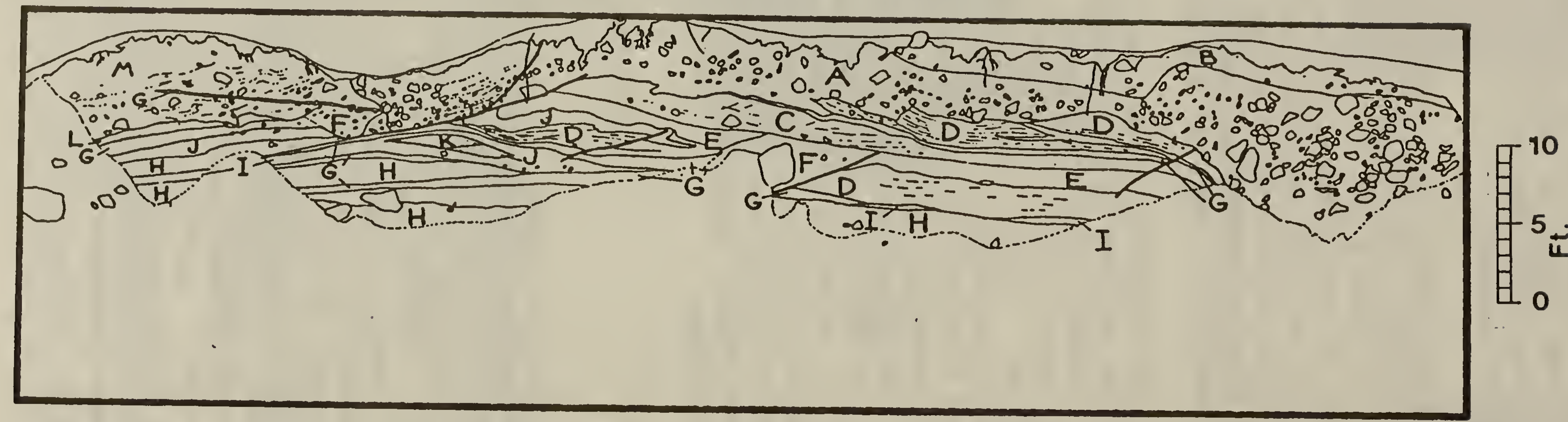
STOP 10 will examine an exposure (see Figure 6) of a kame complex spatially associated with an ice-disintegration moraine in the Moose Pond area of Denmark. My current working hypothesis considers that the "pile" of morainic debris, and the associated kame complex, constitute a single set of deposits that are genetically and temporally related. In this sense, they may constitute a special variation on the theme of morphologic sequences, as defined by Koteff (1974). However, as contrasted with the current interpretation of sequences, which requires the presence of internally active ice to produce the sediment, this interpretation implies that the majority of the stratified sediment has been derived indirectly: either from a debris-laden disintegrating ice mass (which in this case resided in the Moose Pond area), or from the morainic materials themselves. The interpretation does not require the presence of active glacial ice at any point during the formation of the complex. I expect and welcome the inevitable debate concerning these and other similar hummocky morainic deposits in the region.

3.4 MORPHOLOGIC SEQUENCES

Several of the sets of stratified deposits located in the smaller highland basins between the Saco and Ossipee valleys can be considered as morphologic sequences in a strictly descriptive sense. Morphologic sequences, or, in the more euphonious form, "morphosequences" (Koteff, 1980) constitute the surficial mapping equivalents of "rock"-stratigraphic units, in that they can be identified in the field, mapped, and distinguished from other such units on the basis of observable physical properties. Along with the properties conventionally used in hard-rock geology to distinguish rock-stratigraphic units, the geologist working in Quaternary terranes also uses morphology. In fact, morphology is generally a first-order property which is often the only field criteria used in reconnaissance-level surficial mapping projects. In more detailed mapping projects, particularly where abundant topographic data are available, it is often possible to establish morphostratigraphic units on the basis of the surface elevation of deposits, as well as on the basis of the morphological "facies" (ie, esker vs. valley train). The correlation of river terraces is the most obvious example of such an exercise. Similarly, the differentiation of glaciolacustrine deltas on the

FIGURE 6

PICTORIAL SECTION OF SEDIMENTS EXPOSED IN THE MOOSE POND KAME COMPLEX



LEGEND

A Sand boulder gravel, poorly to moderately sorted, crudely stratified; boulders subangular to subrounded.

B Gravelly fine to medium sand, bimodal texture, gravel to small cobble size, massive to crudely stratified.

C Gravelly coarse sand, very well sorted with openwork texture, gravel chiefly pebble-sized, planar bedding.

D Medium sand, very well sorted, planar bedding.

E Silty sandy pebble gravel to silty gravelly sand, moderately well sorted, tabular bedding.

F Gravelly silty sand, poorly sorted, massive diamicton.

G Gravelly silty sand, poorly sorted, matrix-supported, crudely stratified diamicton.

H Gravelly fine sand. Gravel chiefly pebble-sized, crudely laminated. Contains thin (<0.1') interbeds of stratified diamicton similar to unit G.

I Slightly gravelly, silty fine to coarse sand. Moderately well sorted, stratified, gravel chiefly pebble-sized.

J Silty sandy gravel, moderately poorly sorted, gravel clasts from small pebble to large cobble size, clast-supported.

K Pebble gravel, very well sorted with openwork texture, cross-stratification.

L Sandy gravel, moderately well sorted, gravel clasts range from pebble to cobble size.

M Gravelly sand, complexly interbedded with medium sand, silty sand, coarse sand, pebble gravel, and stratified sandy diamicton. Contains clasts to small boulder size.

NOTES

1) Section constructed from photographic mosaics assembled over the course of several years. Because it is a composite of observations gathered as the exposure has developed over time, not all units have been present at all times.

2) Vertical scale as shown. Horizontal scale = vertical in the center of the section only; because of the amphitheater shape of the exposure, there is considerable distortion of scale at edges of the section due to image projection.

the basis of water planes (as constructed from the elevations of their surfaces or their topset/foreset contacts) is another example. In these two cases, a significant change in the elevation of a reconstructed water plane evinces a difference in time. With this in mind it then becomes possible, by means of water-plane correlation procedures analogous to biostratigraphic correlation in bedrock geology, to produce a relative time stratigraphy for a given drainage basin.

STOPS 6, 7 (an optional stop, depending on the quality of the exposure on the day of the trip), **8** and **9** (another optional stop, depending on time) have been included as examples of sets of deposits genetically related to the progressive lowering of the levels of small glacial lakes in north-draining tributary valleys. The various chronological interpretations of these sequences are presented as part of the descriptions for the individual stops. Although the general picture of the depositional setting for these sediments is fairly clear, the details of drainage changes over time, as lower and lower basins became ice-free, are confusing and pose several problems. This is particularly true of the problematic delta system exposed in the Blake pit in Brownfield (**STOP 6**).

The fundamental issue concerning these sequences revolves around the interpretation regarding the activity of ice during their formation. Ideally, the identification of sequences in mapping deglaciation deposits is simply a technique of gathering basic field data. It should not require the prior application of a particular paradigm regarding the physical activity of the associated ice. The deposits at **STOPS 6-9** can therefore be interpreted in two different ways with regard to the activity of ice: one involving downwasting, disintegrating ice; the other involving internally active ice in near proximity. Strictly in terms of the observations from the deposits themselves, neither explanation is in violation of data. However, particularly with respect to **STOP 6**, the first hypothesis involves fewer assumptions, is the simplest, and seems to create fewer problems than it solves. For these reasons, I currently consider it the most valid. I hope that individuals more familiar with the correlation and interpretation of morphosequences than I am will be with us, and that they will consent to lead a discussion.

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FIELD TRIP ROAD LOG

TOTAL TRIP MILEAGE	MILEAGE BETWEEN STOPS	DESCRIPTION
0.00	0.00	Meet at Hiram Falls, overlook, Route 113, Baldwin, Maine.

2.25	2.25	Pass the "Famous Dancemore" hall on the left.
2.45	0.20	Turn right onto Route 117.
3.55	1.10	Cross Saco River.
3.75	0.20	Turn right into Town of Cornish pit.
3.80	0.05	**STOP 1**: Cornish town pit. There are approximately 20' of section exposed here, showing examples of fluvial bedforms. Current directions, as measured at several places in the pit, range from S11W to S85E.
3.85	0.05	Exit pit; turn right onto Route 117. (Note that the road climbs from a lower terrace surface onto the surface exposed at **STOP 1** about 0.1 mi from the pit entrance.)
4.30	0.45	Bear right towards the village of Cornish.
4.60	0.30	Turn right onto Route 25.
4.75	0.15	Turn right next to hardware store.
4.81	0.06	**STOP 2**: Exposure of rhythmically-bedded clay/silt. The composite section exposed here is:

Sand and pebble gravel	(15')
Slightly gravelly medium sand	(7')
Interbedded fine sand, silt, and medium sand; plane-bedded, with soft sediment deformation structures	(9')
Interbedded silty sand and silt with graded bedding	(8')
Rhythmically-Bedded silt and clay/silt; couplets are on the order of 0.1' thick	(11')
Cobble and boulder gravel	(16')

4.87	0.06	Turn left onto Route 25.
6.85	1.98	Pass small pit on right.
7.00	0.15	Cornish-Limington town line.
7.17	0.17	Notice scarp of kame terrace on left.
8.07	0.90	Turn left onto Tucker Road
9.62	1.55	View to left of the Saddleback Hills. Outcrop on right.
9.82	0.20	**STOP 3**: Site of proposed dam for initial stage of Saco valley glacial lake.
11.62	1.80	Bear right through the junction.
12.12	0.50	Four corners. Go straight through.
13.87	1.75	"Ruin Corner". Bear left onto Route 25.
14.86	0.99	North Limington, junction with Route 11. Go straight.
15.64	0.78	Cross Hamlin Brook.
16.64	1.00	Cross Saco River.
16.67	0.03	Turn right onto the River Road.
18.07	1.40	Pass Milt Brown Road on left.
18.57	0.50	Turn left onto Libby Pines Road.
18.64	0.07	Turn right into pit.
18.67	0.03	**STOP 4**: Standish glaciomarine delta. This pit includes an exposure of a topset/foreset contact, whose surveyed elevation is 296' above sea level

(NGVD) (Thompson, Crossen, Borns, and Andersen, 1983). Although the overall quality of the current exposure is not good, the pit has been included as a stop to show the marine sea level indicator nearest the study area. The marine nature of the delta, which at this location is morpholgically a kame, has been established by virtue of an exposure, in the far side of the pit, of Presumpscot Formation clay/silt overlying the granular sediments which constitute the delta foresets.

18.70	0.03	Turn left onto Libby Pines Road.
18.77	0.07	Turn right onto the River Road
22.10	3.33	Turn left onto Route 25; head back toward Cornish.
26.68	4.58	North Limington
28.43	1.75	Pass through "Ruin Corner"
28.78	0.35	Pass Christian Hill Road on left.
30.62	1.84	Cross Back Brook
31.83	1.21	Pass Tucker Road on right.
33.22	1.39	Cornish-Limington town line.
37.87	4.65	Cross Little River in Cornish.
38.27	0.4	Junction of Routes 5 and 25; go straight.
40.52	2.25	Parsonsfield-Cornish town line.
42.36	1.84	Bear left off Route 25 before it crosses the Ossipee River.
45.76	3.4	Take right turn onto Mudgett Road.
46.66	0.9	Right turn onto woods road. Consolidate vehicles here.
47.41	0.75	**STOP 5**: Exposures of "ribbed" moraines.
48.16	0.75	Pick up vehicles. Turn right onto Mudgett Road.
48.56	0.4	Pass pit on left.
49.06	0.5	North Parsonsfield; turn right onto Route 160.
49.46	0.4	Pass elementary school on right.
50.06	0.6	Turn right at Parsonsfield Seminary.
51.96	1.9	Cross Ossipee River.
52.26	0.3	Turn right onto Route 25.
56.86	4.6	Cross over to Route 160, heading north.
57.56	0.7	Cross Ridlon Brook.
57.69	0.13	Turn right at the South Hiram Post Office; continue on 160.
59.09	1.4	Hiram-Porter town line.
61.39	2.3	Enter "The Notch". This large south-southwesterly-trending gorge, which was described by Stone (1899) as having been carved principally by glacial ice, is the local expression of a prominent regional lineament (the bedrock exposed in the gorge is granodiorite, presumably of the Devonian New Hampshire Plutonic Series (Gilman, 1977)). Striations have not yet been found within the notch itself. Although I assume the gorge owes at least <u>some</u> of its current morphology to glacial scour, there is no way of knowing when the scouring occurred. If it took place essentially during the last deglaciation, it could be that the ice retained a considerable degree of internal vigor as it thinned to the point where flow was strongly controlled by very local bed topography. In any event, "The Notch" contained a meltwater system for a

period of time during latest Wisconsin time, and in addition, appears to have served as a spillway for a small proglacial lake.

62.29	0.9	Hiram-Porter town line.
63.69	1.4	Porter-Hiram town line.
66.39	2.7	Pass Burnt Meadow Pond on right.
67.19	0.8	Bear left on Route 160 at the Civil War monument.
67.99	0.8	Bear left at intersection.
69.29	1.3	Turn left at "Patterson's Ponderosa".
70.69	1.4	Turn right just past the Old Blake Farm.
70.89	0.2	Turn left into large, overgrown pit.

STOP 6: Exposure of deltaic sediments.

This pit, which is active on an intermittent basis as a source of road sand for the Town of Brownfield, exposes a section of the foresets of a small delta. The surface slope of the deposit, the direction of dip of the foreset beds, and the position and inclination of meltwater channels on the hillside across the road from the Blake homestead all indicate that the deposit was constructed by a stream, or streams, flowing from the east. The delta is located in the Quint Brook basin, a northward-draining third-order tributary of the Saco.

That the deposit is of glacial origin is assumed, because it is located very close to the drainage basin divide, and I do not believe that the volumes of nonglacial runoff produced from what little basin area remains above the delta could explain the size of the deposit. Presumably, the lake existed because the drainage basin was blocked by ice to the north. For this to be true, an ice margin at the time of sedimentation would by necessity describe a closed loop around the drumlin located 0.5 miles north of the delta, and the lake would be draining over ice, into ice, or along the edge of ice to the north of the drumlin, and out the Shepards River valley towards the Saco. Whether or not such a situation is feasible will surely be a subject of discussion. If the story is essentially correct, it would make a strong case for thinning, stagnating ice in this basin during its deglaciation; if the ice was active and backwasting (which implies that it had a positive surface slope to the north), the lake would be draining <u>into</u> the ice at a point where one would expect the ice to be getting thicker.

71.09	0.20	Back on 'main road'; turn right.
72.19	1.10	Turn right.
72.89	0.70	Cross Cole Brook.
73.19	0.30	Turn right at 'T' junction (New Boston).
74.59	1.40	**STOP 7:** Optional esker exposure.

This is a rapidly evolving exposure, which last year yielded the first unequivocal paleocurrent data from this, the Cole Brook esker system. Those data indicate convincingly that the deposit was constructed from the north, to the south. The bed topography rises to the

south from approximately 580 feet to 830 feet, which is the elevation of the lowest point through which the system that built the esker could exit the Cole Brook basin. There is a lack of collapse features within the core of the esker, which may argue against the idea that the esker was superimposed on the bed from englacial positions. If this interpretation is correct, the the fact that the esker system was constructed in a southerly direction south over north-sloping bed topography indicates that it formed subglacially in a confined tunnel.
The existence of the esker seems to present an argument for relatively thick ice, while the deposits at **STOP 6** seem to imply thin ice. The simplest explanation for this apparent contradiction would state that the esker formed earlier than the delta (and also, by inference, than the other "non-eskerine" deposits in the Cole Brook and Quint Brook basins). This explanation, involving thinning ice, the development of an "esker phase", followed by the development of a "lacustrine phase", is essentially that proposed by Goldthwait and Mickelson (1982) for the deglaciation of the White Mountains.

74.79	0.20	Turn right
76.09	1.30	View to the left (northwest) of Moat and Presidential Ranges in New Hampshire. Note asymmetrical stoss and lee bedrock hills in the foreground, which is all within Maine.
76.49	0.40	Turn left.
76.79	0.30	Pass "Patterson's Ponderosa" again. Go straight.
77.59	0.80	Junction with West Brownfield Road. Cross road into pit.

STOP 8: Pit in kame plateau.
Exposed are 24' of chiefly medium to coarse sand, with interbedded pebble and cobble gravel near the top of the section. Bedforms include horizontal bedding contained in tabular sets, and climbing ripple-drift. Paleocurrents range from S55E to S90E. The sediments are judged to have been deposited subaqueously by a single meltwater system initiating approximately 2 miles north of the pit. The stream entered a body of open, ponded water choked with relict, disintegrating ice masses. A set of lateral channels, cut in a stepwise manner on the south wall of the Shepards River valley, are believed to have acted as the spillways for this, and several other small glacial ponds. The water levels in these ponds dropped sequentially as lower and lower drainage outlets were exposed upon continuous thinning of the ice in the Shepards River valley.

79.79	2.20	Junction with Route 160, Brownfield. Go straight.
80.89	1.10	Junction with Route 113. Turn left at store and proceed into pit.
80.92	0.03	STOP 9: Exposure in the Brownfield sand plain behind Wally's General Store. Optional stop.

General stratigraphy in the section:

Interbedded sand and pebble gravel	(2')
Interbedded sand and silt	(1')
Interbedded medium sand and slightly gravelly sand	(8')

Bedforms exposed at various times have included 0.2' thick tabular sets of cross-stratified sand (lower contacts tangential to set boundaries), and apparently cyclically-bedded sand and pebble gravel. Paleocurrent directions indicate flow from the north and northwest. The deposits are interpreted to have been deposited in a shallow subaqueous environment.

Exposures within the same landform near the village of Brownfield indicate that the deposit was formed from two sources simultaneously: from the north in the Saco valley; and from the west in the Shepards valley. There are two alternate hypotheses which can explain the deposit:

1) All of the sediment was carried by streams fed by glacial meltwaters, implying that there were relict, disintegrating ice masses in the Shepards River valley at a time when the Saco corridor, at least into Fryeburg, was largely ice-free.

2) The western part of the landform was constructed as a nonglacial "fan" very quickly after the Shepards River basin became deglaciated. This explanation can accommodate both the Saco and Shepards River valleys becoming ice-free at quite similar times.

80.95	0.03	Turn onto Route 160; cross Route 113.
81.10	0.15	Cross B&M tracks.
82.40	1.30	Cross Saco River.
83.10	0.70	Pass road to Brownfield Bog on left.
83.90	0.80	Brownfield-Denmark town line.
86.80	2.90	Turn left onto Lake Road.
87.40	0.60	Turn right. Continue on road on a traverse. Note the texture of the sediment (diamicton, exclusively).
88.20	0.80	Turn left.
88.70	0.50	Turn right.
88.80	0.10	Turn right.
88.90	0.10	Turn right back onto Lake Road.
89.10	0.20	Turn into pits.
		STOP 10: Moose Pond kame complex. General features of the stop are described in Section 3.3.2 of the introductory text. This stop is the end of the excursion.

GEOLOGY OF THE EASTERN PORTION OF THE WHITE MOUNTAIN BATHOLITH, NEW HAMPSHIRE

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INTRODUCTION

The Jurassic White Mountain batholith is located in northern Grafton and Carroll Counties, New Hampshire. It is a composite of several overlapping centers of felsic magmatism. Individual centers are strikingly defined by ring dikes and can include plutonic, hypabyssal, and volcanic rock types. Pyroclastic rocks, chiefly rhyolitic tuffs and breccias, are preserved within ring dikes as large subsided blocks. Numerous plutons of subaluminous to peralkaline granites provide an areal continuity to the batholith. The geology of the White Mountain batholith has been mapped by Billings (1928), Billings and Williams (1935), Creasy (1974), Henderson and others (1977), Moke (1946), Osberg and others (1978), Smith and others (1939), and Wilson (1969).

The spatial and temporal geometry of the magmatic centers provides a convenient basis for dividing the batholith into an eastern and a western portion. The eastern portion of the White Mountain batholith (Figure 1) as exposed in the North Conway 15' quadrangle has at least four magmatic centers developed about 175 m.y. ago (Creasy and Eby, 1983). Hitchcock (1878) provided the first outline of the geology of the North Conway quadrangle and certain of his formational names are still in use, e.g. the Conway Granite. However, the work of Billings (1928) remains the chief geological reference for the North Conway quadrangle. Creasy (1974; unpublished maps), Davie (1975), Osberg and others (1978), and Parnell (1975) have provided greater detail to Billings' pioneering work.

GEOLOGY OF THE NORTH CONWAY QUADRANGLE

The geology of the North Conway quadrangle (Figure 1) is summarized in terms of the major magmatic and structural units of the White Mountain batholith.

The Mt. Osceola Granite, a green amphibole ± biotite granite, is the <u>oldest</u> member of the White Mountain magma series exposed in the North Conway quadrangle (Osberg and others, 1978). The number and original extent of plutons of the Mt. Osceola Granite within the North Conway quadrangle is not fully certain due to the complexity and abundance of younger rocks. A whole-rock Rb-Sr isochron for samples from both eastern and western portions of the batholith yields an age of 186 m.y. (Eby and Creasy, 1983) and indicates synchronous intrusion over a broad area . [This age places the Mt. Osceola as the <u>youngest</u> member associated with the large magmatic center that forms the western portion of the batholith.]

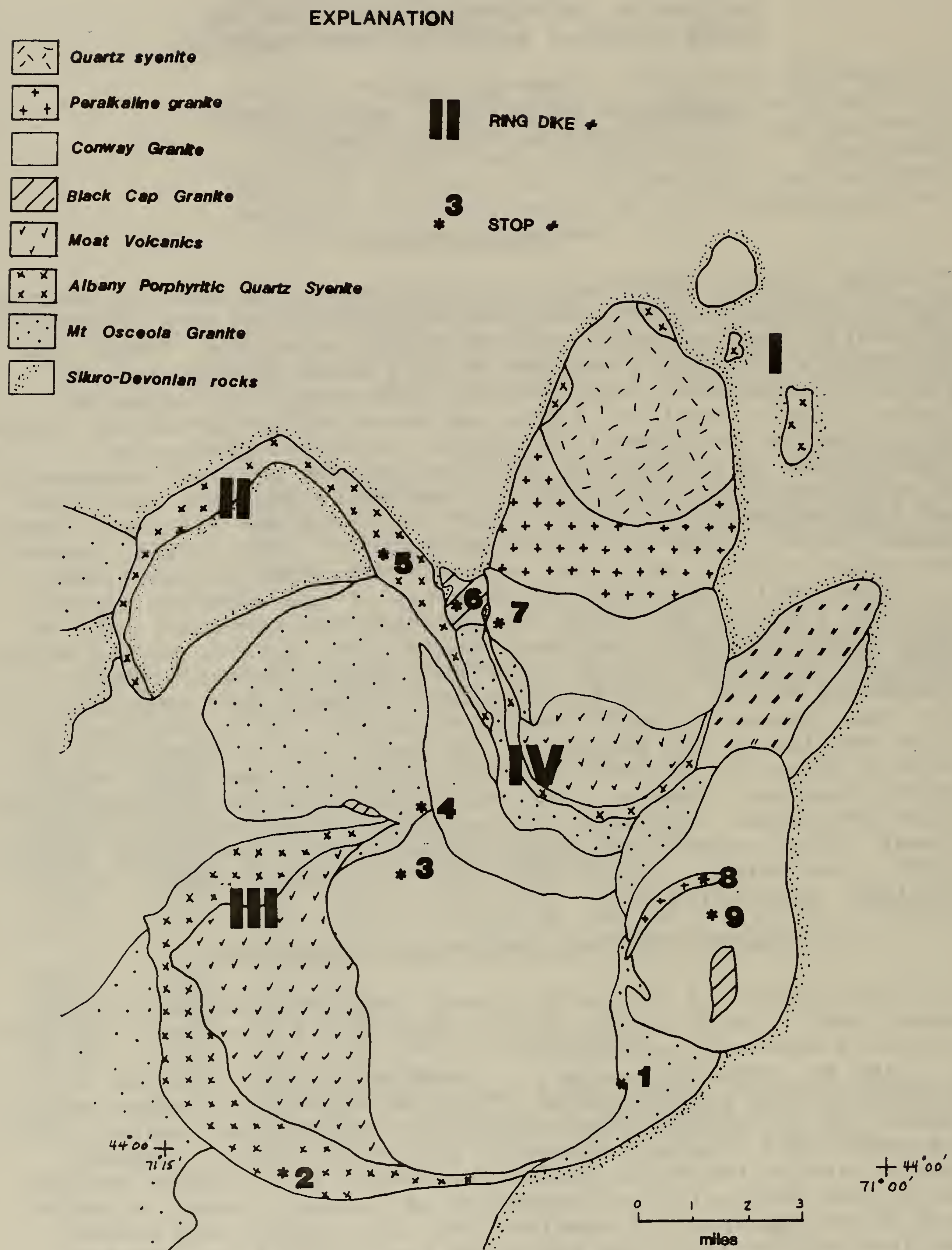

Figure 1. Generalized geologic map of the North Conway quadrangle, NH (after Billings, 1928; Osberg and others, 1978).

At least four magmatic centers are defined by ring dikes of the Albany Porphyritic Quartz Syenite (Figure 1). A whole rock Rb-Sr isochron for samples from three ring dikes yields an age of 175 m.y. (Eby and Creasy, 1983). The ring dikes are not seen to cut each other but relationships to other units indicate their emplacement was not simultaneous. These ring dikes are outwardly dipping at 40°-80° with well developed chill margins adjacent to older rocks. In fine structure these ring dikes are themselves multiple intrusions. At least four separate and petrologically distinctive intrusions of Albany occur in ring III (Figure 1 and e.g. STOP 2). Inclusions present in ring II document a similar structural relationship (e.g. STOP 5).

The Moat Volcanics (Billings, 1928) are exposed in the Moat Range to the west of North Conway; a second major occurrence is on Mount Kearsarge to the northeast. About 3 km of volcanic stratigraphy are exposed in the Moat Range. This section of comendite tuffs and subordinate breccia and trachyte shows pronounced and laterally persistent layering (Billings, 1928; Noble and Billings, 1967) striking northwest and dipping 30°-40° to the northeast. The eutaxitic structure of tuffs yields similar orientations. On geochemical grounds Creasy and Eby (1981) recognize at least four alkali rhyolites in the Moat Range. A whole-rock Rb-Sr isochron for the Moat Volcanics yields an age of 175 m.y. (Eby and Creasy, 1983); K-Ar dates of 168-170 m.y. (Eby, 1986, pers. comm.) are in general agreement. On Mount Kearsarge and on South Moat Mtn the volcanic rocks are intruded by the Albany PQS. Hence the Moat Volcanics are considered as structural blocks dropped down along younger arcuate fractures now largely occupied by the Albany PQS (Billings, 1928; Osberg and others, 1978). However, their thickness and distribution would be consistent with syn- or post-subsidence accumulation (Noble and Billings, 1967). The eruptive source and original extent of the volcanics is a major question: the volcanic lithologies preclude close genetic association with the ring dikes.

The Conway Granite, most extensive unit in the North Conway quadrangle, is a pink biotite granite. Billings (1928) showed the Conway Granite as a single irregularly shaped pluton and as the youngest of the White Mountain magma series. More detailed mapping (Osberg and others, 1978) has recognized several distinct plutons of biotite granite on the basis of texture and outcrop geometry (Figure 1). Absolute ages for plutons in the North Conway quadrangle are within the range of 175 ± 6 m.y. (Eby and Creasy, 1983). Field relations suggest that emplacement of these plutons was not synchronous across the quadrangle but related to individual magmatic centers. The Birch Hill pluton (Osberg and others, 1978) is the largest pluton. The Conway Granite of this pluton becomes finer grained, porphyritic, and miarolitic where it intrudes the Moat Volcanics (e.g. Stop 3). The Gardiner Brook pluton (Osberg and others, 1978) intrudes Moat Volcanics on Mount Kearsarge and is associated with the magmatic center defined by ring IV (Figure 1). The Conway Granite of this pluton shatters Silurian metasediments along the East Branch of the Saco River (STOP 7). A third pluton underlies most the Green Hills, the

prominent north-south oriented ridge forming the east side of Mt. Washington Valley; this pluton is well exposed on Black Cap mountain (STOP 9).

The Black Cap Granite (Billings, 1928) is a fine-grained pink biotite granite that outcrops in two small areas in the North Conway quadrangle (e.g. Stop 6). It is shattered and intruded by the Conway Granite (Green Hills pluton) on the flanks of Black Cap (Stop 9). Billings (1928) considered this rock an early lithologically distinct `phase' of the Conway Granite. Osberg and others (1978) suggest that the Black Cap granite may be coeval with and a roof facies of the Conway Granite.

Peralkaline granite forms an arcuate dike and small intrusion within the Conway Granite of the Green Hills pluton (Stop 8). A K-Ar date of 177 ± 7 m.y. (Eby, 1984, pers. comm.) is obtained from this locality. Riebeckite granite forms large areas of outcrop (e.g. on North Doublehead) that are seemingly young plutons spatially associated with hastingsite granite (Mt. Osceola Granite ?). The lack of clear contact relations and the presence of riebeckite + hastingsite granites suggest a genetic association as well.

DESCRIPTION OF UNITS (Table 1 and Figure 1)

The <u>Mt. Osceola Granite</u> is a medium- to coarse-grained hypersolvus granite that is dark green where fresh. It consists of an interlocking network of anhedral to subhedral microperthite 3-10 mm in diameter enclosing rounded grains of smokey quartz. Ferrohastingsite and locally annite are interstitial late crystallized minerals. Fayalite and ferrohedenbergite are frequently present in minor amounts and encased by reaction rims of ferrohastingsite. Characteristic accessories include allanite, sphene, zircon, fluorite, and monazite. Locally the Mt. Osceola is peralkaline with ferrorichterite or riebeckite present rimming ferrohastingsite.

The <u>Moat Volcanics</u> consist of comendite, trachyte, and tuff-breccia. The comendites are crystal-rich lithic tuffs; eutaxitic texture typical of welded ash flow tuffs is present within the section exposed in the Moat Range. These blue-gray to pink rocks contain variably abundant phenocrysts (1-3 mm) of quartz and sanidine (now microperthite) and rare phenocrysts of biotite, ferrohastingsite, ferrohedenbergite, and riebeckite set in a matrix of quartz and alkali feldspar. Accessories include apatite, fluorite, zircon, and magnetite. Lithic fragments constitute 1-5% of the comendite and range from a few cm to a few mm in size. Lithic types include hornfels, cogenetic volcanic rocks, porphyritic quartz syenite, and rarely cogenetic plutonic rocks. The trachyte consists of phenocrysts (2-3 mm) of pink alkali feldspar set in a dense (<0.1 mm) groundmass of alkali feldspar. Accessories include abundant hematite and minor zircon, magnetite, epidote and clinozoisite. The tuff-breccia contains angular to subrounded blocks ranging from a few cm to a m in size. These are generally polymict breccias but locally may be dominated by a single lithology. Lithic fragments include a

variety of metamorphic rock lithologies, Paleozoic intrusive rocks, and cogenetic volcanic and hypabyssal rocks. The red to gray matrix is generally sub-microscopic; where resolved it is quartz and alkali feldspar with chlorite, sericite, biotite, magnetite, and hematite.

The Albany Porphyritic Quartz Syenite contains phenocrysts of microperthite (5-10 mm) and quartz (2-4 mm) in all occurrences the abundance of phenocrysts and of quartz:feldspar varies. This variation is noted both within a ring dike (e.g. #2a,b,c of Table 1) and among different ring dikes. Minor phenocryst phases include ferrohedenbergite, fayalite, and ilmenite. The groundmass consists of anhedral quartz, alkali feldspar, and ferrohastingsite; the latter also forms poikilitic reaction rims on the phenocrysts of mafic minerals. The groundmass is relatively uniform in grain size (<2-3 mm) within an intrusion (except near contacts) but shows variation among different intrusions. Accessories include allanite, sphene, zircon, and fluorite.

The Conway Granite is a medium- to coarse-grained pink biotite two-feldspar granite. Values of microperthite:oligioclase range from 2:1 to 10:1 and average 4-5:1. Biotite forms anhedral interstitial grains up to 5 mm in size. In contrast with other members of the White Mountain magma series, fayalite and ferrohedenbergite are not present. Field definition of the Conway Granite excludes samples containing amphibole in hand sample (Osberg and others, 1978). Zircon, allanite, apatite, and sphene are common accessories. Near contacts the Conway Granite shows a variety of textures that may grade into each other on the outcrop scale: porphyritic, aplitic, miarolitic, and pegmatitic. Weakly developed banding on the cm- to dm-scale resulting from variations in grain size and/or mineral concentrations is developed near some contacts. Lithic fragments of any type are sparse in the Conway Granite.

The Black Cap Granite is a fine-grained (0.1 mm) equigranular rock composed of quartz, microperthite, subordinate oligoclase, and chloritized biotite. Accessories include zircon, magnetite, apatite, and fluorite.

The peralkaline granite is composed of subhedral grains of white microperthite (5-10 mm) and clear quartz (2-6 mm), blocky interstitial grains of riebeckite-arfvedsonite (<10 mm), and flakes and aggregates of interstitial biotite. Characteristic of this rock are abundant radiating arrays of golden colored astrophyllite. Fluorite, ilmenite, sphene, and apatite are common accessory minerals. Near contacts, miarolitic pods and cavities are developed on a cm-scale; here prismatic riebeckite crystals are found upto 5 cm in length.

Table 1. <u>Modes of plutonic rocks from the North Conway quadrangle listed according to stop number [Figure 1]</u> (Billings, 1928; Osberg and others, 1978; and Davie, 1975).

<u>Stop No.</u>	#1a	#1b	#2a	#2b	#2c
Quartz	35	22	10	16	20
Alkali feldspar	49	72	79	70	68
Plagioclase	14	3	0	0	0
Biotite	2	tr	1	0	tr
Amphibole	0	1	8	9	
Ferrohedenbergite	0	tr	8	2	1
Fayalite	0	2	1	tr	tr
Accessories	tr	tr	3	3	3

<u>Stop No.</u>	#5	#6	#7	#8	#9
Quartz	13	29	33	39	28
Alkali feldspar	74	48	49	54	47
Plagioclase	0	17	16	0	18
Biotite	1	6	2	1	7
Amphibole	12	0	0	5	0
Ferrohedenbergite	0	0	0	0	0
Fayalite	0	0	0	0	0
Accessories	tr	tr	tr	1	1

1a Conway Granite, Birch Hill pluton, Hurricane Mtn Road.
1b Mt. Osceola Granite, Rattlesnake Mtn, Redstone area.
2a,b,c Albany Porphyritic Quartz Syenite, three distinct types present within ring dike III, Little Attitash Mtn.
5 Albany Porphyritic Quartz Syenite, Jackson Falls.
6 Black Cap Granite, Thorn Mtn.
7 Conway Granite, Gardiner Brook pluton, Burnt Knoll Brk.
8 Riebeckite granite, North Doublehead.
9 Conway Granite, Green Hills pluton, Black Cap Mtn.

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ITINERARY

Assembly point is the parking lot of Burger King restaurant located at the junction of Routes 16 and 302 between Conway and North Conway villages, NH. Assembly time is 7:45 a.m. Late arrivals: we will pause momentarily at the assembly point following Stop 1 at about 9:15 a.m.; we should be at Stop 2 until about 10:15 a.m.

All stops are located in the North Conway 15' topographic quadrangle and on Figure 1. Access to Stops 2 and 3 is through the Ossipee Lake 15' quadrangle.

Mileage

0.0 exit parking lot using easterly driveway to Route 302; turn right (east) onto Route 302.

(0.6)

0.6 Turn left (north) at Redstone town triangle; Redstone quarry prominently exposed on south end of Green Hills.

(0.1)

0.7 Cross tracks of Maine Central Railroad

(0.1)

0.8 Pass gated dirt road on left (south) and Mason house on right.

(0.1)

0.9 Park where paved road turns left (south) and woods road continues straight and uphill; proceed on foot along woods road about 0.1 mi to Stop 1a.

STOP 1a, b: Redstone Quarry: Conway Granite (1a) and Mt. Osceola Granite (1b). The coarse pink Conway Granite (Table 1, #1a) of the Birch Hill pluton is homogeneous in grain size and texture and was quarried extensively for building stone. Sparse inclusions, chiefly of porphyritic granite, are present locally. A few thin (3 cm) dikes of aplitic granite are seen on the main bench of the quarry. Samples from here have been used in numerous studies related to the high concentrations of U and Th in the Conway Granite (e.g. Adams and others, 1962; Birch and others, 1968; Osberg and others, 1978).

return to cars and walk .1 mi south on paved road to gated dirt road. The dirt road ends in a small clearing (.1 mi); note the capped drill hole. About

3000 ft of core were pulled from this hole to evaluate the geothermal heat potential of the Conway Granite (DOE's hot dry rock program). The site is on the contact between the Mt. Osceola and Conway Granites; both granites (as well as Albany PQS) are present in the core. Exit the clearing adjacent to columnar pieces of granite. Follow path .1 mi to small quarry with derricks (Stop 1b).
The Mt. Osceola Granite (Table 1, #1b) was quarried here. The contact between this green ferrohastingsite granite and the Conway Granite of Stop 1a is reached by traversing around the east and north (top) sides of the quarry along a narrow foot trail. The Conway Granite is distinctly finer grained at the contact.

Retrace route to cars; return to junction of Routes 16 and 302, reset odometers.

@@@@@@@@

0.0 Junction Routes 16 and 302; turn left (south) and continue to Conway village.

(3.0)

3.0 Turn right (west) on Route 113 and continue through Conway village.

(0.8)

3.8 Turn right (west) on Route 112, the Kancamagus Highway.

(0.8)

4.6 Cross Albany town line.

(1.5)

STOPS 2a, b, c: Swift River: The Kancamagus Highway crosses ring dike III (Figure 1) at a low angle providing excellent outcrops of the Albany Porphyritic Quartz Syenite. At least three Albany-type lithologies (Table 1, #2a,b,c) are present and are distinguished by the abundance and proportion of quartz and feldspar phenocrysts. Davie (1975) demonstrated the composite nature of ring dike III to the north (Attitash Ski Area) but the exposures are less accessible. This trip will stop at three places along the Kancamagus highway to illustrate these differing lithologies. At all three locations, pull cars as far off pavement as possible.

6.1 STOP 2a

(0.4)

6.5 Enter White Mountain National Forest

(2.0)

8.5 Cross Hobb's Brook

(0.9)

9.4 STOP 2b

(0.3)

9.7 STOP 2c

(0.6)

10.3 Turn sharply right (east) on Dugway Road.

(0.2)

10.5 Cross Albany covered bridge; Albany PQS (type locality of Hitchcock?) crops out here.

(4.1)

14.6 Dugway Picnic Area on right (south) side of road.

(0.8)

15.4 South Moat Trail sign on left (north) side of road. South Moat Mtn on skyline to north.

(0.9)

16.5 Red Eagle Pond along left (north) side of road.

(1.6)

18.1 Junction with West Side Road; turn left (north) onto West Side Road; reset odometers.

@@@@@@@@@

0.0 Junction of West Side Road (=Dugway Road) with Still Road, proceed north on West Side Road.

(1.3)

1.3 Cross tracks of Boston and Maine Railroad.

(1.3)

2.6 Unmarked dirt road on left (west) side of road leads to USFS Smokey Quartz Collecting Area. [This is an accumulation of boulders of the miarolitic contact facies of the Conway Granite; bedrock is not exposed]

(1.0)

3.6 Birch Hill Road enters on left (west); the quarry at the summit has provided samples of Conway Granite for geochemical and geochronologic studies.

(1.6)

5.2 Entrance to Echo Lake State Park on the left (west); continue straight ahead.

(0.6)

5.8 Stop sign at T intersection; turn left (north) and continue on West Side Road. [A right turn here will take you to north end of North Conway village]

(0.6)

6.4 Cathedral Ledge Road on left (west); continue on West Side Road. [Side road leads to summit of Cathedral Ledge (3 mi round trip) which provides excellent views of the Saco River and Mt. Washington Valley.]

(0.7)

7.1 Turn left (west) onto dirt road just north of the Lucy farm on right; park here. This locality is reached by a 1.8 mi traverse (3.6 mi round trip) entirely on graded road and trail; there is 400 ft of relief, up going and down returning.

STOP 3: Diana's Baths and Lucy Brook: Moat Volcanics and Conway Granite of the Birch Hill pluton in contact. At Diana's Baths (0.4 mi) the coarse Conway Granite becomes porphyritic and contains large (upto 3 m) segregations of dark material that Billings (1928) interpreted as partially assimilated inclusions. NO HAMMERS PLEASE!

From here, follow the Moat Mtn Trail (1.4 mi) to the first trailside ledges (Moat Volcanics), then cut over to Lucy Brook (100 ft).

The contact between the Moat Volcanics (upstream) and the Conway Granite (downstream) is very well exposed here in Lucy Brook. In the lowest stream exposures, the coarse Conway Granite grades into medium-grained porphyritic contact facies characterized by a heterogeneous and miarolitic texture. Large rounded inclusions of fine-grained biotite granite are included in the Conway Granite (as at Stop 9), perhaps representing an older contact facies. Blue-gray comendite of the Moat Volcanics is bleached and pink where cut by numerous fractures and thin (<2 cm) quartz veins. Eutaxitic texture is not well developed here; the rocks resemble quartz porphyry.

return to vehicles and continue north on the West Side Road.

(1.8)

8.9 Park on wide left shoulder of road.

STOP 4: Humphrey's Ledge: Mt. Osceola Granite is exposed at the north end of Humphrey's Ledge. This is an excellent locality for collecting samples that contain fayalite and ferrohedenbergite.

return to vehicles and continue north on West Side Road.

(2.6)

11.5 Stop sign. Turn right (northeast) on Route 302.

(2.3)

13.8 Turn left onto Route 16 at Glen village; Use caution--this is a very dangerous intersection! reset odometers.

@@@@@@@@@

0.0 Junction of Routes 16 and 302 at Glen village; proceed north on Route 16.

(2.4)

2.4 Bear right on Route 16A crossing Jackson covered bridge; proceed through Jackson village past Yesterday's Restaurant (on right), bear left (west) past small triangular "park" and cross stone bridge. [note location of Thorn Hill Road and Thorn Mtn Road for later reference.]

(0.7)

3.1 Turn right (north) on Route 16B after crossing stone bridge and proceed uphill; Albany PQS exposed to right.

(0.3)

3.4 Park on wide right (east) shoulder at Jackson Falls picnic area. Walk to broad exposures in Wildcat Brook.

NO HAMMERS PLEASE!

STOP 5: Jackson Falls: Albany Porphyritic Quartz Syenite in a composite ring dike. This is one of the best localities to examine the Albany PQS--about 2000 ft of continuous exposure is present between here and Jackson along Wildcat Brook. Just downstream of the iron bridge a screen of Siluro-Devonian gneisses and granite 110 ft wide is intruded by the Albany PQS. Downstream from this screen, the Albany is relatively uniform in mineralogy and texture and contains a small proportion of small (2-5 cm) inclusions. Upstream, large (upto 1 m) inclusions of feldspar-poorer Albany-type lithology enclosed by the Albany PQS are an earlier intrusion within ring dike II. A wide (1 m) dike of fine-grained pink biotite granite cuts the Albany PQS and screen of country rocks. This may be related to the Black Cap Granite seen at Stop 6.

return to vehicles and continue North (uphill) on Route 16B.

(0.1)

3.5 Turn right and continue across Wildcat Brook.

(0.3)

3.8 Turn right (south, downhill) on this branch of Route 16B and return to Jackson village.

(0.6)

4.4 Bear left at Route 16A then turn immediate left (north) onto Thorn Mtn Road; continue steeply up passing side roads to left and right.

(1.5)

5.9 Continue straight (uphill) through uncontrolled 4-way intersection.

(0.7)

6.6 T-junction at saddle of Middle Mtn (left) and Thorn Mtn (right); turn left (west) then immediate right onto wide dirt road passing base lodge of defunct Tyrol Ski Area.

(0.1)

6.7 Park left of 2-bay garage. A 1 mi loop at Stop 7 traverses ski trails to and from the summit of Thorn Mtn.

STOP 6: Thorn Mtn: Black Cap Granite, Albany PQS, and pre-White Mountain igneous and metamorphic rocks. The ski trails on the north slope of Thorn Mtn expose pavement outcrops of the fine-grained and homogeneous Black Cap biotite granite (Table 1, #6). Just below (north of) the summit, the Black Cap intrudes a screen of country rocks consisting of Siluro-Devonian metamorphic rocks intruded by Devonian (?) granites and pegmatites. The south margin of the screen is shattered and intruded by a chill facies of Albany PQS. The Albany at and south of the summit has coarser grained

groundmass and may be a separate intrusion. The screen is cut out to the east; the Black Cap Granite is in contact with the chilled Albany but age relations are not evident.

return to cars and retrace route to Jackson village.

(2.2)

8.8 Junction with Route 16A; bear left and reset odometers.

@@@@@@@@@

0.0 Junction of Thorn Mtn Road and Route 16A; proceed south on Route 16A.

(0.2)

0.2 Turn left onto Thorn Hill Road.

(3.1)

3.3 Bear left (south) at junction with Route 16A across from Swiss Chalets Motel.

(0.5)

3.8 Turn left (north) onto Town Hall Road at 4-way intersection (no stop sign) and proceed along the East Branch of the Saco River.

(1.9)

5.7 Park on wide right shoulder at south end of bridge; proceed to Stop 7 across the bridge.

STOP 7: Gardiner Brook: Conway Granite of the Gardiner Brook pluton. The coarse-grained Conway (Table 1, #7) exposed at the north end of the long outcrop becomes porphyritic towards the south where Siluro-Devonian gneisses occur. These country rocks are shattered and intruded by a fine-grained biotite granite (Black Cap Granite ?) which is intruded by the porphyritic Conway Granite.

return to cars and retrace route to Route 16A.

(1.9)

7.6 Turn left (south) onto Route 16A; proceed to junction with Route 302.

(1.9)

9.5 Junction with Route 302 at Intervale. Turn left (south) on Route 302.

(0.1)

9.6 Turn left (east) onto Hurricane Mtn Road; reset odometers.

@@@@@@@@@

0.0 Junction Route 302 and Hurricane Mtn Road, Intervale; proceed east on Hurricane Mtn Road.

(1.6)

1.6 Kearsarge Mountain Trail sign on left (north); continue straight.

(2.3)

3.9 Height-of-land on Hurricane Mtn Road; pull off on right side of road; park here for Stops 8 and 9.

About 50 ft east of the height-of-land a woods road leaves the north side of the road; follow this to a point where a trail leaves left (about .25 mi) and ascends to open ledges (.15 mi) of Stop 8.

STOP 8: Hurricane Mtn. Riebeckite ± biotite granite is exposed in the open south-facing ledges at this stop; Conway Granite is exposed around the rest of the mountain. Riebeckite granite (Table 1, #8) intrudes the Conway Granite in the Green Hills to the south. The blast debris and ledges at Stop 8 contain abundant miarolitic pods and cavities notable for their large prismatic crystals of riebeckite-arfvedsonite. This miarolitic contact facies is probably the top or margin of a larger subjacent pluton of peralkaline granite (c.f. South Doublehead Mtn). Exposures near the summit of Hurricane Mtn lack biotite and contain only scattered miarolitic pods and may lie below this contact zone. Astrophyllite is a characteristic accessory mineral in these peralkaline rocks.

return to the Hurricane Mtn Road and walk 50 ft east towards the height-of-land to a sign on the south side of the road marking the beginning of the trail to Black Cap Mtn. Follow trail to the open ledges of Black Cap Mtn (1 mi).

STOP 9: Black Cap Mtn. Conway Granite of the Green Hills pluton is a medium-grained pink biotite granite (Table 1, #9). The granite is well exposed in the ledges as the trail ascends the north slope of Black Cap. Near and south of the summit, the granite locally becomes sub-porphyritic. About .25 mi southeast of the summit (no trail) the fine-grained Black Cap Granite is exposed in a geometry that suggests a sub-horizontal sheet. The Black Cap Granite is shattered and intruded by the Conway Granite of Stop 8. The Black Cap may represent a marginal roof facies of the Green Hills pluton preceding the main intrusive pulse of Conway Granite.

return to the Hurricane Mtn Road and the cars.

END To Lewiston: Continue east on the Hurricane Mtn Road; at the first and only stop sign (T-junction) turn right (south) onto the road to East Conway. This road joins Route 113 in about 3 miles. Continue south on Route 113 to junction with Route 302 in Fryeburg, ME. Follow Route 302 east through Fryeburg, Bridgton, and Naples. Turn north on Route 11 south about 2 mi east of Naples and follow to Auburn. Consult detailed map of Lewiston-Auburn area provided with registration materials for route to Bates College.

TRIP B-3

STRATIGRAPHY AND METAMORPHISM OF THE SILURIAN AND LOWER DEVONIAN ROCKS OF THE WESTERN PART OF THE MERRIMACK SYNCLINORIUM, PINKHAM NOTCH AREA, EAST-CENTRAL NEW HAMPSHIRE

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ACKNOWLEDGMENT

At the outset we would like to point out the fundamental and critical role of Professor M. P. Billings in deciphering the geology of the Pinkham Notch area. Although we were not able to find a copy of his road log for it, Billings led an NEIGC field trip to the Pinkham Notch area in 1946, and it is probable that some of the exposures to be visited on the present trip were included on Billings' trip of 40 years ago. Our map of the area differs from those of Billings primarily in the names and ages of some of the units. The differences are readily explained, however, by the fact that the work in Maine, which developed the Merrimack synclinorium stratigraphic sequence and tied many of the units to fossils, was not done until years after Billings' maps of the Pinkham Notch area were published. Billings' study of the stratigraphy, structure, and metamorphism of the Pinkham Notch area is a classic upon which this field trip is thoroughly dependent.

GEOLOGIC SETTING

The Merrimack synclinorium forms a broad belt of Silurian and Lower Devonian strata east of the Bronson Hill-Boundary Mountain anticlinorium across eastern Maine, central New Hampshire, and central Massachusetts and Connecticut (Williams, 1978; Osberg and others, 1985; Billings, 1956; Zen and others, 1983; Rodgers, 1985) (fig. 1). Moench (1971) defined and mapped a distinctive sequence of stratigraphic units in the west part of the synclinorium in the Rangeley area of western Maine. Hatch and others (1983) extended this Rangeley area sequence southwest across eastern and south-central New Hampshire. Thompson (1984), Robinson (1981), and Berry (1985), among others, have extended all or parts of this sequence across southernmost New Hampshire and central Massachusetts into northern Connecticut. Eusden and others (1986) have reported this same sequence of units further east in the synclinorium in southeastern New Hampshire and adjacent southwestern Maine. These Silurian and Lower Devonian strata are interpreted to have been deposited at or near the eastern margin of continental North America in the time interval after the Taconian collision between North America and the Bronson Hill island arc and before the Acadian collisional (?) orogeny.

The Silurian and Lower Devonian rocks of this sequence in east-central New Hampshire are divided into the formations indicated in table 1. All are metamorphosed sedimentary rocks; no metavolcanic rocks have been recognized in the belt in eastern New Hampshire and they are very rare or absent throughout the belt.

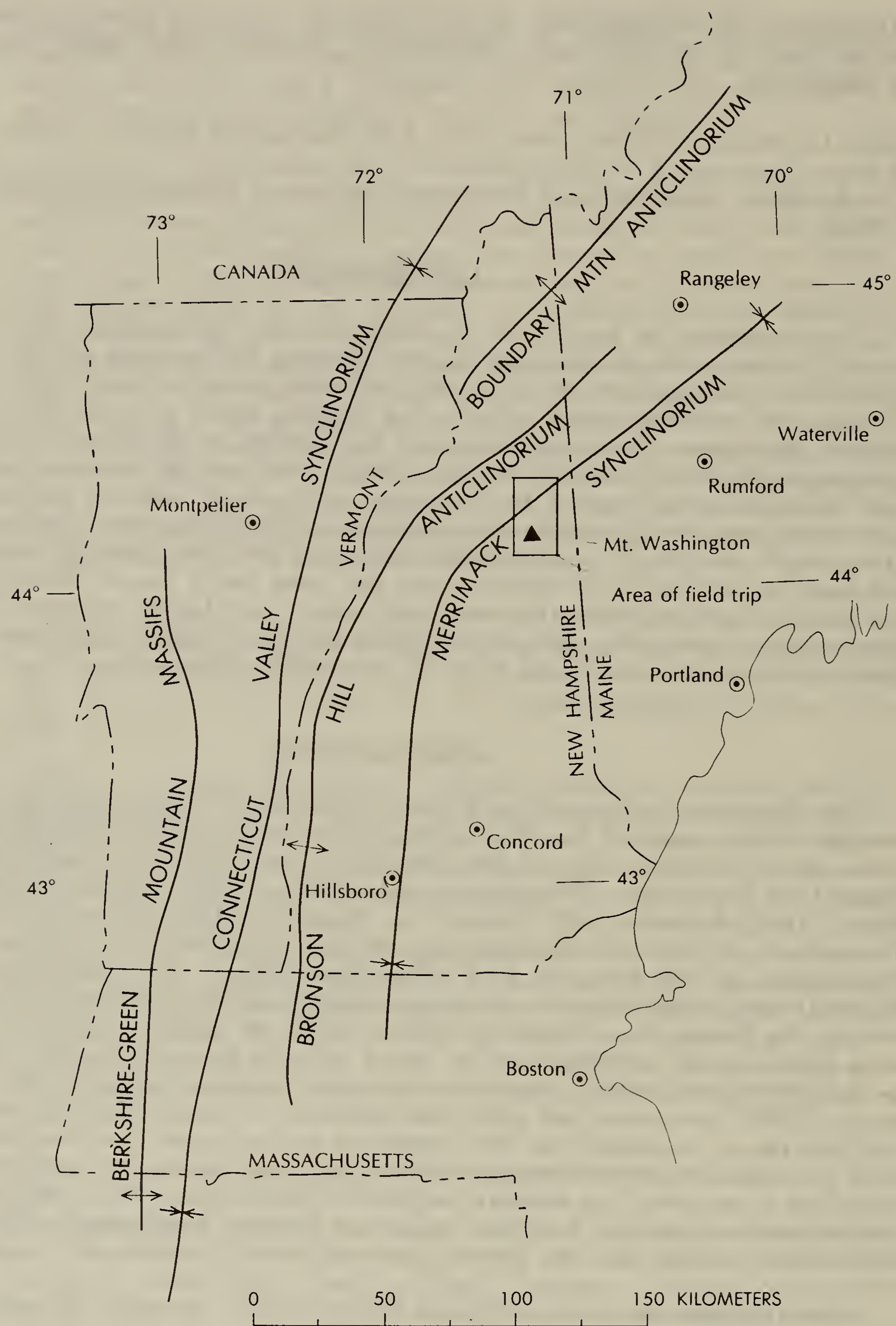


Figure 1. Map showing location of the field trip area in the geologic framework of western New England.

TABLE 1. STRATIGRAPHIC UNITS OF THE MERRIMACK SYNCLINORIUM IN EAST-CENTRAL NEW HAMPSHIRE.

AGE	FORMATION	LITHOLOGY
Early Devonian	Littleton	Mica schist and granulite
Silurian	Madrid, upper	Plagioclase-quartz-biotite granulite
	Madrid, lower	Layered calc-silicate granulite
	Smalls Falls	Sulfidic schist and quartzite
	Perry Mtn.	Quartzite and mica schist
	Rangeley	Mica schist, quartzite, grit

The model advanced by Moench and others (1982) and by Hatch and others (1983) is that the Silurian rocks of the Merrimack synclinorium are an eastern, more distal, deeper water facies of the shelf-facies Silurian Clough Quartzite and Fitch Formation of the Bronson Hill anticlinorium (figs. 1 and 2). The axis of the Bronson Hill anticlinorium is about 15 miles west or northwest of the area to be visited on this field trip. The Silurian sequence to be seen on the trip is intermediate in both thickness and number of units between the Silurian of the Bronson Hill (Billings, 1937, 1956) and of the Merrimack synclinorium east of Rangeley, Maine (Moench, 1971). The on-strike continuation of the Lower Devonian Littleton Formation in north-central Maine, the Seboomook Formation, has been interpreted by Hall and others (1976) to have been derived from the east in contrast to the underlying westerly-derived Silurian rocks. The Littleton forms a thick blanket across both the shelf (Bronson Hill) and basin (Merrimack) Silurian sequences in New Hampshire (fig. 2).

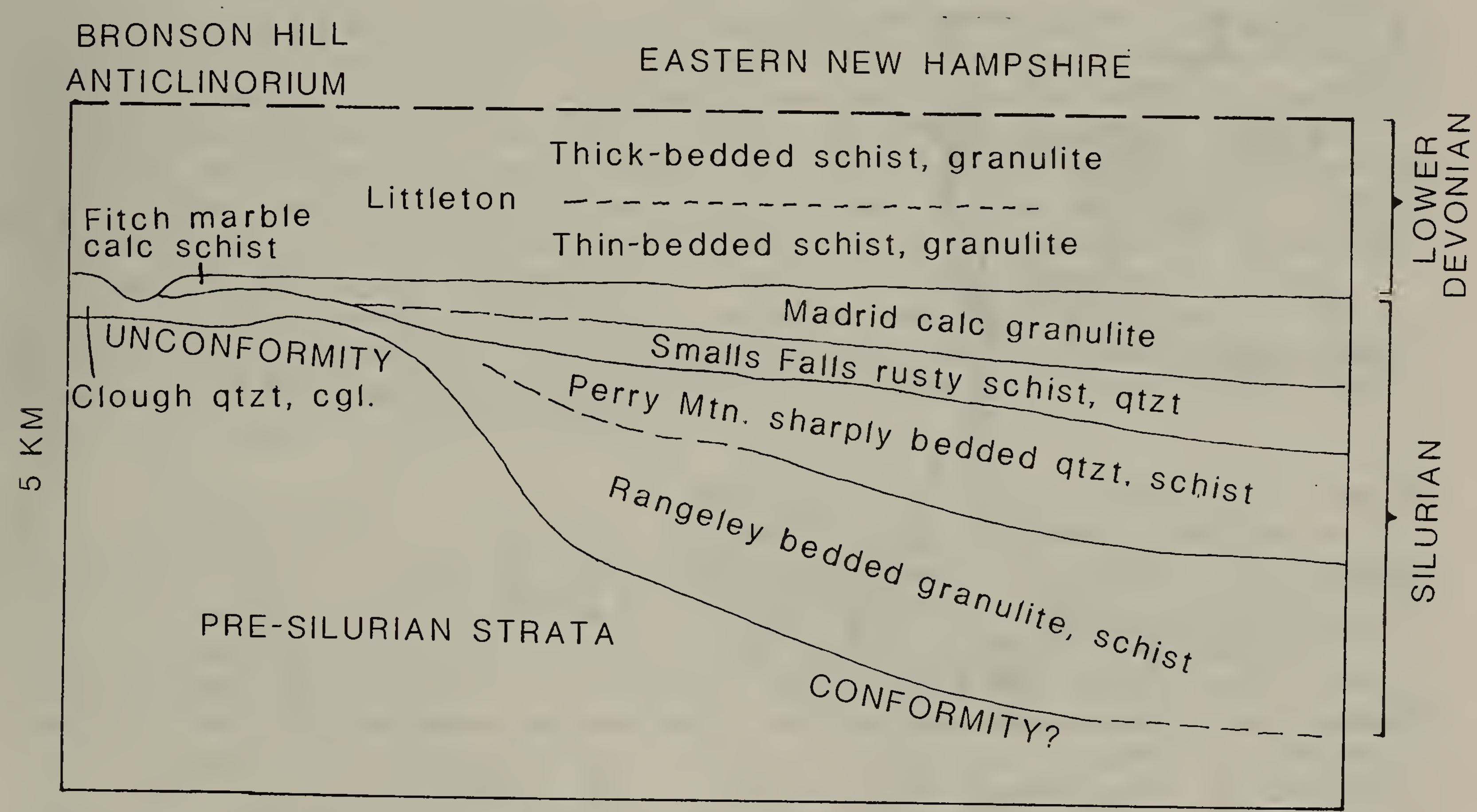


Figure 2. Stratigraphic diagram showing the relation between the thin Silurian shelf facies on the Bronson Hill anticlinorium and the much thicker Silurian basin facies of the Merrimack synclinorium to the east. Both are blanketed by roughly comparable thicknesses of easterly derived (?) Lower Devonian Littleton Formation.

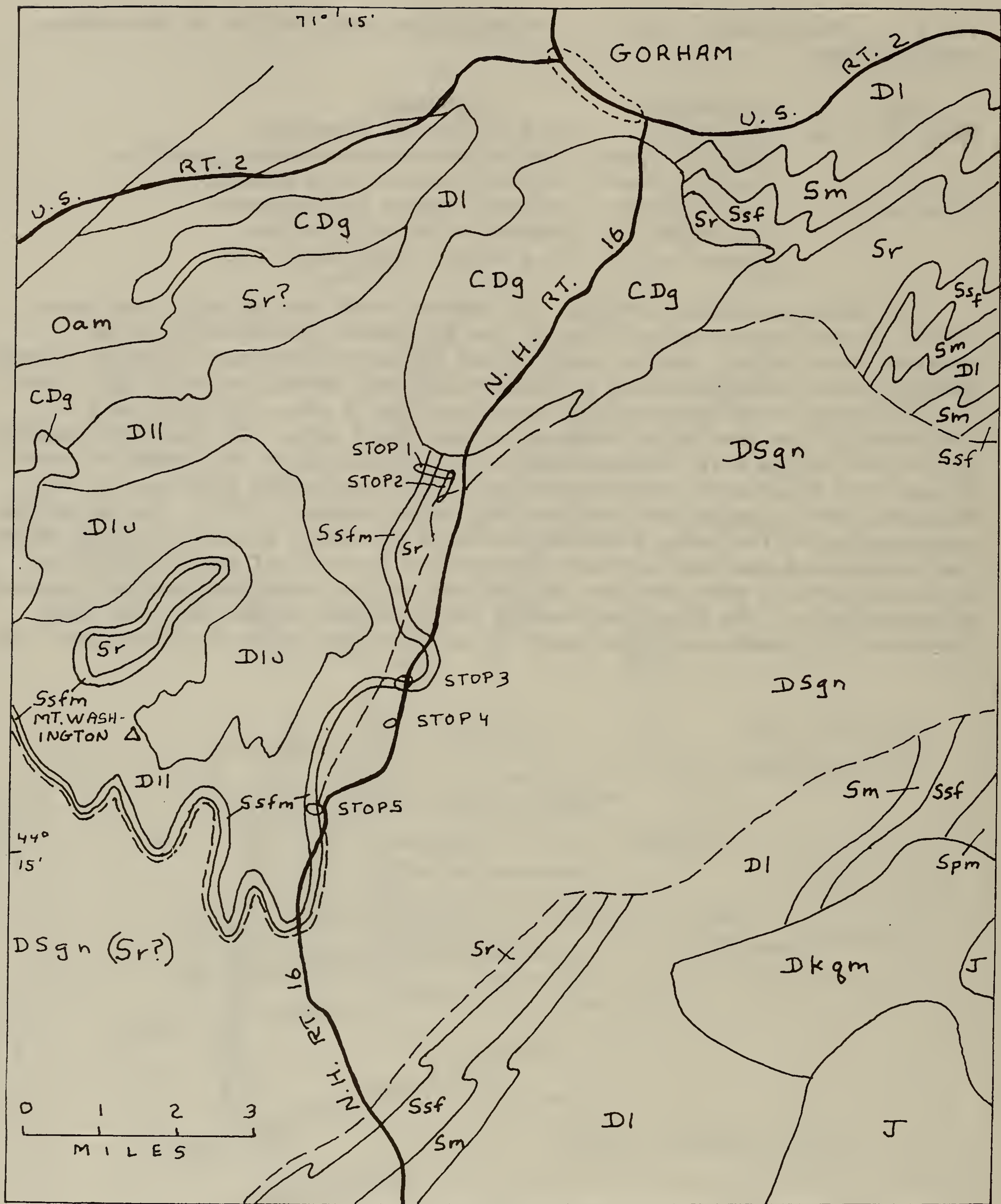


Figure 3. Geologic map of Pinkham Notch and the surrounding area. Modified from Billings and others (1946), Billings and Fowler-Billings (1975), Billings (1928), and Hatch and Moench (1984).

EXPLANATION FOR FIGURE 3

J	Jurassic to Cretaceous plutonic rocks
CDg	Devonian to Carboniferous granitic rocks
Dkqm	Devonian Kinsman Quartz Monzonite
DSgn	Undifferentiated Silurian and Devonian paragneiss
Dl	Undivided Lower Devonian Littleton Formation
Dlu	Upper member of the Littleton Formation
Dll	Lower member of the Littleton Formation
Ssfm	Undivided Smalls Falls and Madrid Formations
Sm	Silurian Madrid Formation
Ssf	Silurian Smalls Falls Formation
Spm	Silurian Perry Mountain Formation
Sr	Silurian Rangeley Formation
Oam	Ordovician Ammonoosuc Volcanics
	Contact, approximately located
	Approximate location of gradational boundary between undifferentiated paragneiss and recognizable formations

The field trip will examine exposures in the southwest part of the old Gorham 15-minute quadrangle (Billings and Fowler-Billings, 1975) and the southeast corner of the Mount Washington 15-minute quadrangle (Billings, 1941; Billings and others, 1946) (fig. 3). More modern topographic maps of these areas are the Carter Dome 7.5-minute quadrangle (southwest quarter of the Gorham 15-minute), and the Mount Washington 7.5- X 15-minute quadrangle (south half of the Mount Washington 15-minute).

The rocks to be seen on the trip include exposures of the Silurian Rangeley, Smalls Falls, and Madrid Formations and the Lower Devonian Littleton Formation (table 1, fig. 3). The Perry Mountain Formation was not recognized in the immediate area of the trip, although it is present north of Route 2 northeast of Gorham. The Perry Mountain is also present about 8 miles east of the trip area on Eagle Crag of the Baldface Range in the southeast part of the Gorham 15-minute quadrangle (south-central edge of the Wild River 7.5-minute quadrangle) (east edge of figure 3). There it is characterized by gray, sharply bedded, thin-bedded (2-8 cm) quartzite and pelitic schist. Hatch believes that the absence of the Perry Mountain from the Pinkham Notch area results from nondeposition, perhaps because of proximity to the basin margin (fig. 2), rather than from faulting.

All of the rocks to be seen on the trip, here mapped as Rangeley, Smalls Falls, Madrid, and Littleton, were included in the Littleton Formation by Billings (1956), but some of them had been assigned to different stratigraphic units in earlier interpretations by Billings (table 2). The first modern detailed map of the rocks in the general vicinity of Mount Washington and Pinkham Notch was prepared by Billings (1941). This map, which covered the southern half of the Mount Washington 15-minute quadrangle plus adjacent parts of the Gorham and Crawford Notch 15-minute quadrangles, assigned the rocks we will look at to the Partridge, Fitch, and Littleton Formations. On this map the schists and gneisses of the Ordovician Partridge and Lower Devonian Littleton Formations were separated from each other by a very narrow belt of calc-silicate rocks assigned to the Silurian Fitch Formation. Five years later, Billings and others (1946) published a map of the entire Mount

Washington 15-minute quadrangle, which included most of the area of the 1941 map. On this map, and in the accompanying report, all of the schists and gneisses previously assigned to the Partridge Formation were reassigned to the Littleton Formation, and the thin calc-silicate unit previously assigned to the Fitch was named the Boott Member of the Littleton from exposures on Boott Spur on the south ridge of Mount Washington. Thus, all of the stratified rocks were reinterpreted to be within the Lower Devonian Littleton Formation, and only the Boott was separately mapped. This same stratigraphic system was followed by Billings (1956) on the statewide report and map and by Billings and Fowler-Billings (1975) on their map of the Gorham 15-minute quadrangle.

TABLE 2 STRATIGRAPHIC ASSIGNMENTS OF THE ROCKS IN THE PINKHAM NOTCH AREA SHOWING CHANGES IN INTERPRETATION, NOMENCLATURE, AND AGE THROUGH TIME.

Billings (1941)		Billings et al.,(1946)		This report	
Littleton	(Dev)	Littleton	(Dev)	Littleton	(Dev)
Fitch	(Sil)	Boott Member	(Dev)	Madrid	(Sil)
				Smalls Falls	(Sil)
Partridge	(Ord)	Littleton	(Dev)	Rangeley	(Sil)

The rocks Billings and others (1946) assigned to the Boott are a very distinctive light- and dark-green layered calc-silicate granulite not reported from the Littleton in the Bronson Hill anticlinorium (Billings, 1937, 1956). Many students of New England geology have pondered at length, although little has been written, over whether the Boott was really a calc-silicate member stratigraphically within the Littleton, or was Fitch Formation below the Littleton as originally assigned, or possibly neither. During a quick trip to exposures on the West Branch of the Peabody River (figs. 6, 10) in the fall of 1977, Robert Moench first introduced Hatch to the possibility that the Boott might be the Madrid Formation of the Rangeley, Maine, sequence. Subsequent re-study of the metamorphic rocks of the southern part of the Mount Washington and Gorham 15-minute quadrangles and of the northern part of the Crawford Notch and North Conway 15-minute quadrangles supports this interpretation. The strongly layered green calc-silicate granulite (fig. 4) is identical in thickness, bedding style, and composition to the lower part of the Madrid Formation of western Maine (Moench, 1971). Furthermore, at virtually every locality of this unit in eastern New Hampshire it adjoins deeply rusted, pyrrhotite-rich, graphitic, flaggy quartzite and schist typical of the Smalls Falls Formation, which underlies the Madrid in western Maine. These graphitic and sulfidic rocks may have influenced Billings in his original (1941) assignment of the rocks below the calc-silicate unit to the Partridge Formation.

The lithology of the rocks on both sides of the layered calc-silicate and sulfidic schist units is also compatible with the correlation with the Rangeley, Maine sequence. The layered calc-silicate granulite grades, over about 10 meters, into thick-bedded to massive plagioclase-quartz-biotite, "salt and pepper" granulite with local minor calc-silicate beds and pods. This salt and pepper granulite in turn grades into well-bedded, light-gray aluminous pelitic schist and granulite. Within about 10 meters of the contact, the aluminous schist and granulite (quartzite) unit is characterized by cyclically graded beds 5 to 10 cm thick that consistently indicate that the sequence is topping up from the pyrrhotitic rocks (Smalls Falls) into the layered calc-silicate (lower Madrid) into the massive plagioclase-quartz-

Figure 4. Photograph of the well-layered calc-silicate granulite of the lower part of the Madrid Formation. West Branch of the Peabody River.

Figure 5. Photograph of well-bedded schist and granulite of the Rangeley Formation with calc-silicate "football". Stop 2A, Peabody River.

biotite "salt and pepper" granulite (upper Madrid), and finally into the aluminous schist and granulite with graded bedding (Littleton).

Stratigraphically below the deeply rusted pyrrhotitic rocks of the Smalls Falls are generally well bedded schist and quartzite that are broadly similar to the aluminous schist and granulite of the Littleton Formation. These rocks differ from the Littleton rocks, however, in the following subtle ways. 1) The rocks below the Smalls Falls contain abundant pods, generally 15 to 30 cm thick and 30 cm to 1 m long, of calc-silicate granulite (fig. 5). In order to side-step the question of whether these pods are concretions or boudins, they are commonly referred to as "footballs". 2) Local beds preserve primary grading and indicate that this unit underlies the Smalls Falls. More typically, however, the progression from granulite to schist is abrupt on both sides of a given granulite bed, and primary tops are indeterminate. The Littleton Formation has a significantly higher percentage of graded beds. 3) Many exposures of this unit are slightly rusty weathered, and commonly this rustiness has a distinctive reddish cast unlike that in either the Smalls Falls or the Littleton. 4) The granulites of this unit are locally more quartz rich than those of the Littleton. 5) Locally the granulites are gritty or pebbly. Because all of these features are characteristic of the well-bedded schist and granulite parts of the Rangeley Formation in western Maine, we have made that correlation (Hatch and others, 1983; Hatch and Moench, 1984). The only argument against this correlation is the local absence in the Pinkham Notch area of the Perry Mountain Formation, which in Maine occurs between the Smalls Falls and the Rangeley. As noted above, we feel this absence is due to nondeposition of the Perry Mountain in the western part of the depositional basin near the (Bronson Hill) shelf.

Present stratigraphic thicknesses in the Pinkham Notch area can be accurately measured for the Smalls Falls and Madrid Formations but can only be roughly estimated for the Rangeley and Littleton Formations. Because bedding is no longer recognizable in most of the Rangeley Formation in the Pinkham Notch area, its internal structure is difficult or impossible to determine. For this reason the stratigraphic thickness can only be estimated from the large areas of exposure to be at least a kilometer. In contrast, the Smalls Falls Formation can be accurately measured to range from about 6 meters thick at Lakes of the Clouds, just south of the summit of Mount Washington, to possibly 100 meters thick near Pinkham Notch. The lower, layered calc-silicate part of the Madrid Formation similarly can be accurately measured to be only about 10 meters thick. The upper, plagioclase-quartz-biotite granulite part of the Madrid ranges up to a few hundred meters in thickness, although its gradational contact with the overlying Littleton makes precise measurement difficult. The Littleton was reported by Billings (1941, p. 895) to be about 4000 feet (1220 meters) thick in the area immediately north of Mount Washington, and we agree with that figure.

Ages assigned to the units in the Pinkham Notch area are all based on fossil control from outside the area (Hatch and others, 1983). The Silurian age of the Rangeley derives from late Llandoverian fossils from the Blanchard Ponds belt of the formation in western Maine northwest of Rangeley (Moench and Boudette, 1970). The age of the Smalls Falls depends upon correlation with the fossiliferous Parkman Hill of central Maine (Pankiwskyj and others, 1976). No fossils have been reported from the Madrid, and its Silurian age is dependent upon correlation with the latest Silurian (Pridolian) (Harris and

others, 1983) Fitch Formation in the Bronson Hill anticlinorium. The Early Devonian age of the Littleton Formation is based on fossils from the type area (Billings and Cleaves, 1934).

METAMORPHISM

All of the rocks to be seen on the trip have been metamorphosed to sillimanite grade. A small area immediately southwest of the Dolly Copp campground is staurolite-andalusite grade (fig. 6; Billings, 1941, p. 888) but will not be seen on the trip. The most striking difference in the sillimanite-grade rocks is textural, between schist and granulite sequences that preserve primary compositional layering (bedding), to be seen at Stops 1 and 2A, and comparable rocks of the same formation that have been changed to gneiss and in which bedding has been obliterated, to be seen at Stops 2D and 4. Of interest is the fact that the two types coexist in very close proximity across a moderately sharp boundary. On Stop 2 we will see exposures that show the transition between the two types.

As shown by both Billings (1941, fig. 3) and by Hatch and Moench (1984), the gneiss-schist metamorphic boundary approximately, but not precisely, follows the stratigraphic contact between the Rangeley and Smalls Falls Formations in the field trip area (fig. 3). Regardless of the orientation of that stratigraphic boundary surface at the time of metamorphism, it is difficult to believe that the closely coincident gneiss-schist boundary surface is truly isogradic. Rather, it seems probable that the close geographic coincidence of these two boundaries, which are unrelated in origin, must result at least in part from a greater inherent propensity of the Rangeley schist, relative to the Littleton schist, to be converted into gneiss. This propensity may be a function of subtle differences between the chemical composition of the Rangeley schist and that of the outwardly similar Littleton schist, but this question has yet to be studied in detail. The reluctance of the Smalls Falls and lower Madrid rocks to be converted to gneiss will be shown at Stop 3.

Although irregular in detail, the boundary separating the gneissic rocks from bedded schists and granulites roughly parallels the regional trend of isograds through this part of central New England, where the grade increases eastward from a low in the Connecticut valley into easternmost New Hampshire and adjacent Maine. Taken by themselves, these relations suggest a prograde metamorphic control on the transformation to gneiss. In the vicinity of the field trip stops, the gneiss boundary drawn on figure 3 is, although gradational over a few hundred meters, a fairly sharp boundary. The rocks to the west, around Mount Washington, are well-bedded metasedimentary rocks, whereas the rocks to the east of Route 16 are gneisses. The gneiss boundary shown in the northeast and southeast parts of figure 3 is much less precise. It essentially bounds areas in which enough exposures of the distinctive Madrid and Smalls Falls lithologies were present to enable mapping them and assigning the adjoining rocks to either the Rangeley or the Littleton, from areas in which this was not possible. Most of the rocks presumed to be either Rangeley or Littleton on both sides of the boundary are gneisses. Although too few exposures of distinctive Madrid or Smalls Falls rocks were seen in the large area of DSgn on figure 3 to enable mapping them, most of the gneisses contained calc-silicate "footballs", or had the distinctive red-rusty cast, or both, suggestive of the Rangeley Formation. This interpretation is

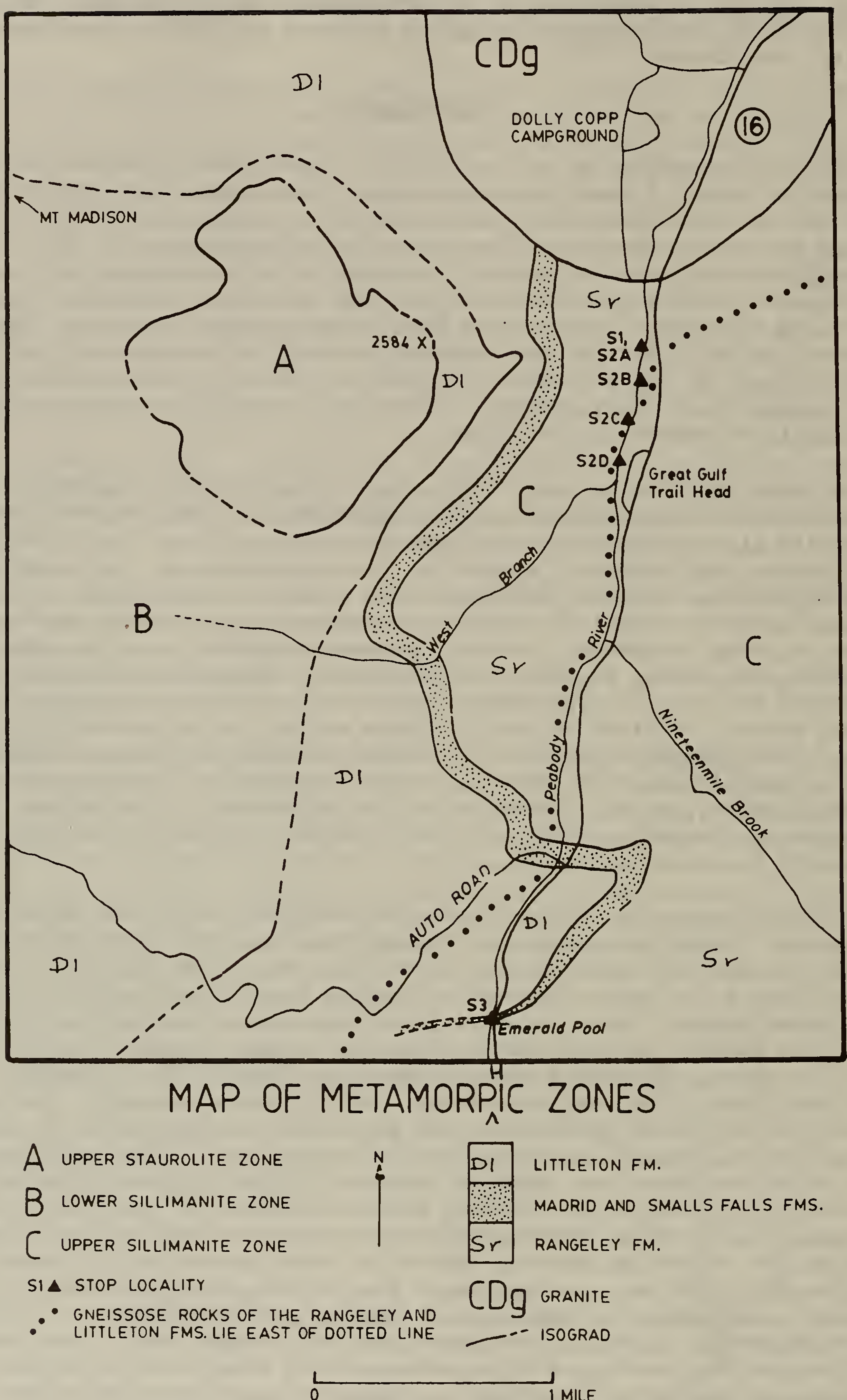


Figure 6. Map showing metamorphic zones in the vicinity of Stops 1 through 3.

significantly different from that of Billings and Fowler-Billings (1975) in which three separate pre-"Boott" calc-silicate units extend northeasterly across the area of DSgn on figure 3. Although we are very uncertain about their continuity across the area of DSgn, most of the exposures of these calc-silicate rocks seen by Hatch impressed him as being lower Madrid.

A detailed study of the metamorphism in a 12 square mile area roughly centered around the 2584 foot knob on the eastern slope of Mt. Madison (figs. 6 and 10) reveals evidence for polymetamorphism in the Pinkham Notch area (Wall and Guidotti, 1986). The first metamorphic event, M1, is shown by the presence of abundant pseudomorphs of muscovite, quartz, and sillimanite after andalusite. The second event, M2, was at sillimanite grade and formed the pseudomorphs. Three metamorphic zones are mapped in this area: an upper staurolite zone, a lower sillimanite zone, and an upper sillimanite zone (fig. 6). Figure 7 shows the AFM and the AKNa topologies indicating the observed assemblages in each of the three metamorphic zones.

The boundary between the upper staurolite and the lower sillimanite zones is defined by the appearance of the sillimanite + biotite join in the AFM topology for the lower sillimanite zone. In many metapelites elsewhere, sillimanite is brought in by a discontinuous reaction: Staur + Chl + NaMusc + Qtz $\rightleftharpoons$ Bio + Sill + K-richer Musc + Ab + H_2O. This specific tie line flip is not observed in the present study because the transition to the lower sillimanite zone occurs within the Littleton Formation, which has a relatively iron-rich bulk composition (compared to the more Mg-rich Rangeley). Hence, primary chlorite is absent. Had the transition to the lower sillimanite zone occurred in the Rangeley Formation (as it does in western Maine), the tie line flip probably could have been documented in terms of observed assemblages. Because of fairly abundant Fe-sulfides, the "silicate bulk composition" of the Rangeley Formation is relatively richer in Mg, thereby enabling the occurrence of Mg-rich phases like chlorite.

Within the lower sillimanite zone, a systematic decrease in the modal percent of staurolite is observed as the upper boundary of this zone is approached. The complete disappearance of staurolite by the reaction Staur + NaMusc + Qtz $\rightleftharpoons$ Bio + Sill + K-richer Musc + Ab + Gar + H_2O (Guidotti, 1970) defines the boundary between the lower and upper sillimanite zones. The breakdown of staurolite results in the formation of the three-phase field Sill + Bio + Gar on the AFM topology for the upper sillimanite zone (fig. 7). Mineral assemblages observed in the upper sillimanite grade schists include Sill + Bio + Gar, Bio + Gar, and Sill + Bio. In the Littleton Formation all of the observed assemblages include ilmenite $\pm$ graphite. In the Rangeley Formation, pyrrhotite is present in addition to ilmenite and graphite.

The metamorphism that produced the present pattern of zones (M2) overprints an earlier metamorphism of at least staurolite + andalusite + biotite grade. The evidence for the earlier metamorphic event (M1) lies in the numerous euhedral prograde pseudomorphs after staurolite and andalusite present in the pelitic schists of the Rangeley and Littleton Formations throughout the area. The pseudomorphs were formed during M2. Both of these events are static in nature and are considered to be Acadian in age, but the recent report by Lux and Guidotti (1985) of Carboniferous metamorphism in western Maine invites speculation on this point.

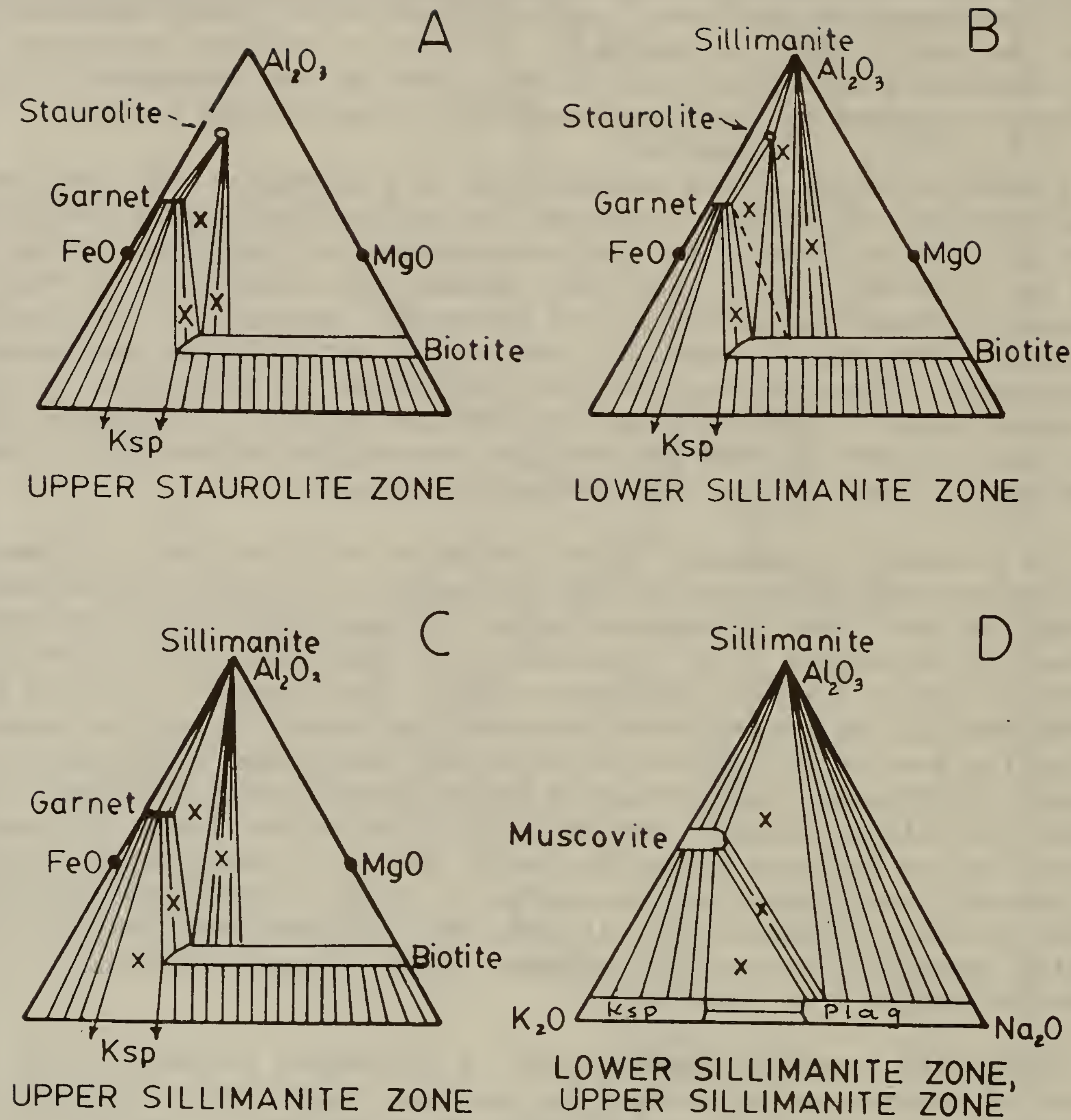


Figure 7. Schematic AFM (A,B,C) and AKNa (D) projections of assemblages (observed assemblages shown by x) for the upper staurolite through upper sillimanite zones. Diagrams A, B, and C correspond to the assemblages observed in areas marked A, B, and C respectively on figure 6.

The progressive transition from well-bedded Rangeley schist and granulite to Rangeley gneiss will be seen on Stop 2. All of the exposures examined on Stop 2 are within the upper sillimanite zone (fig. 6). At Stop 2A bedding, some with grading, is still well preserved. At Stops 2B, 2C, and 2D, over a distance of less than a kilometer south of Stop 2A along the Peabody River and essentially parallel to the strike of bedding, the well-bedded schists and granulites of the Rangeley Formation gradually change to moderately rusty-weathering, foliated, spangled gneisses (fig. 8). This textural change occurs over an even shorter distance (about 500 meters) east of Stop 2A across Route 16. In the first stage of this process the bedding becomes less well defined and grading is no longer recognizable. Pseudomorphs become less abundant and some muscovite spangles as much as 1 cm in diameter are present (Stop 2C, fig. 8). Further south up the Peabody River at Stop 2D, the rock becomes distinctly gneissose with alternating thin layers 0.5 to 2 cm thick of lighter

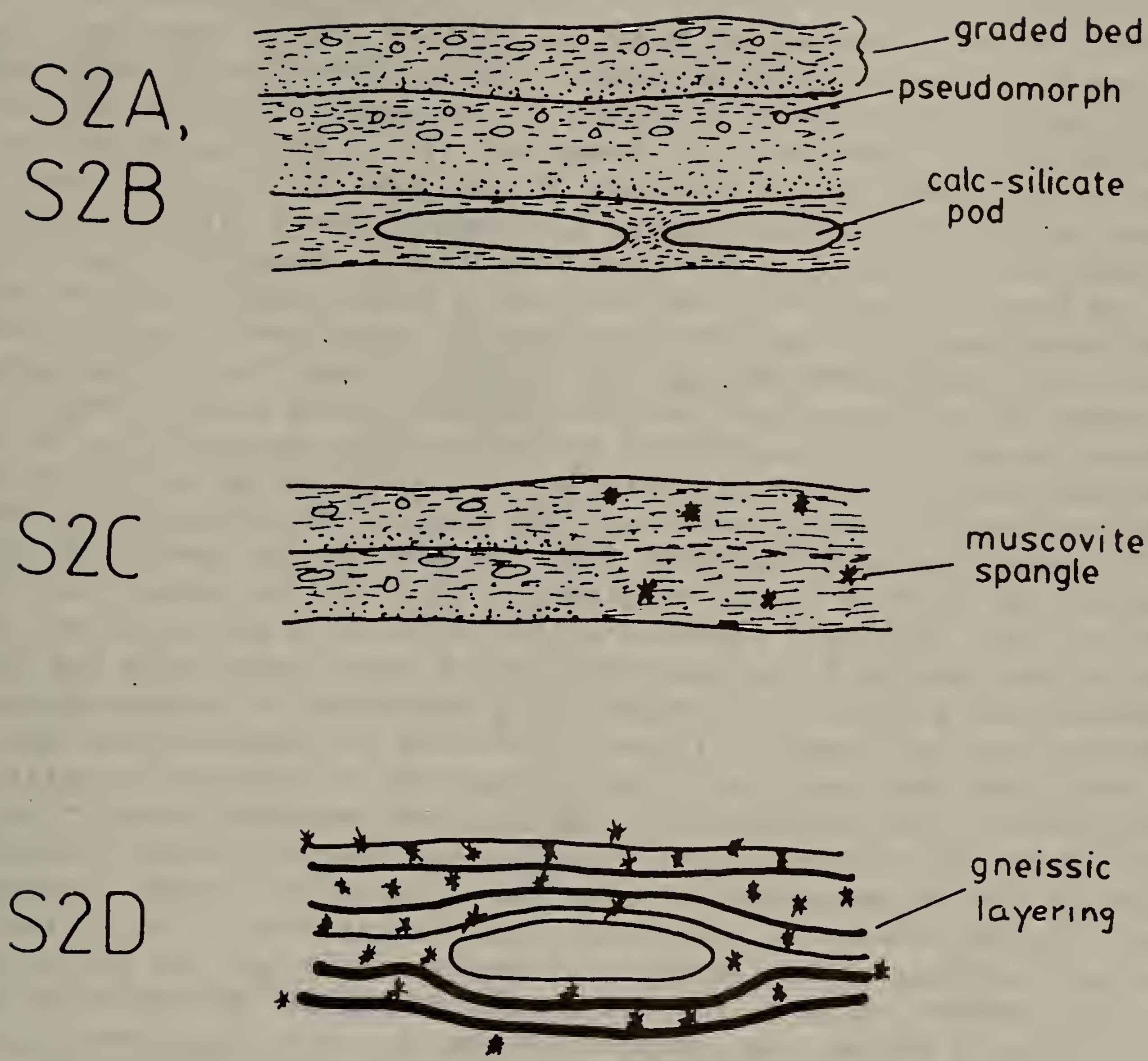


Figure 8. Schematic diagram illustrating the transition from well-bedded schist and granulite to gneiss in the Rangeley Formation at Stop 2. See text for discussion.

quartzofeldspathic and darker biotite-rich material, suggestive of incipient anatexis (Stop 2D on fig. 8). No hint of the original bedding remains, and the pseudomorphs have disappeared. Clots of coarse-grained quartz, plagioclase, and muscovite 10 to 15 cm in length are oriented both parallel and transverse to the foliation, which is defined by parallel biotite plates. Muscovite spangles as much as 2 cm across are more abundant in these gneissose rocks. Calc-silicate pods, present in horizons parallel to bedding in the bedded Rangeley schists and granulites, persist in the gneissose rocks where they are generally oriented parallel to the foliation.

Available outcrop data indicate that the gneiss exposed along Route 16 between the Mt. Washington Auto Road and Emerald Pool is Littleton Formation stratigraphically above the Madrid Formation. This gneiss differs from that of the Rangelely Formation in that calc-silicate pods ("footballs") are rare to absent, muscovite spangles up to 8 cm in length appear to have replaced muscovite pseudomorphs after andalusite, and its color is gray rather than rusty red-brown.

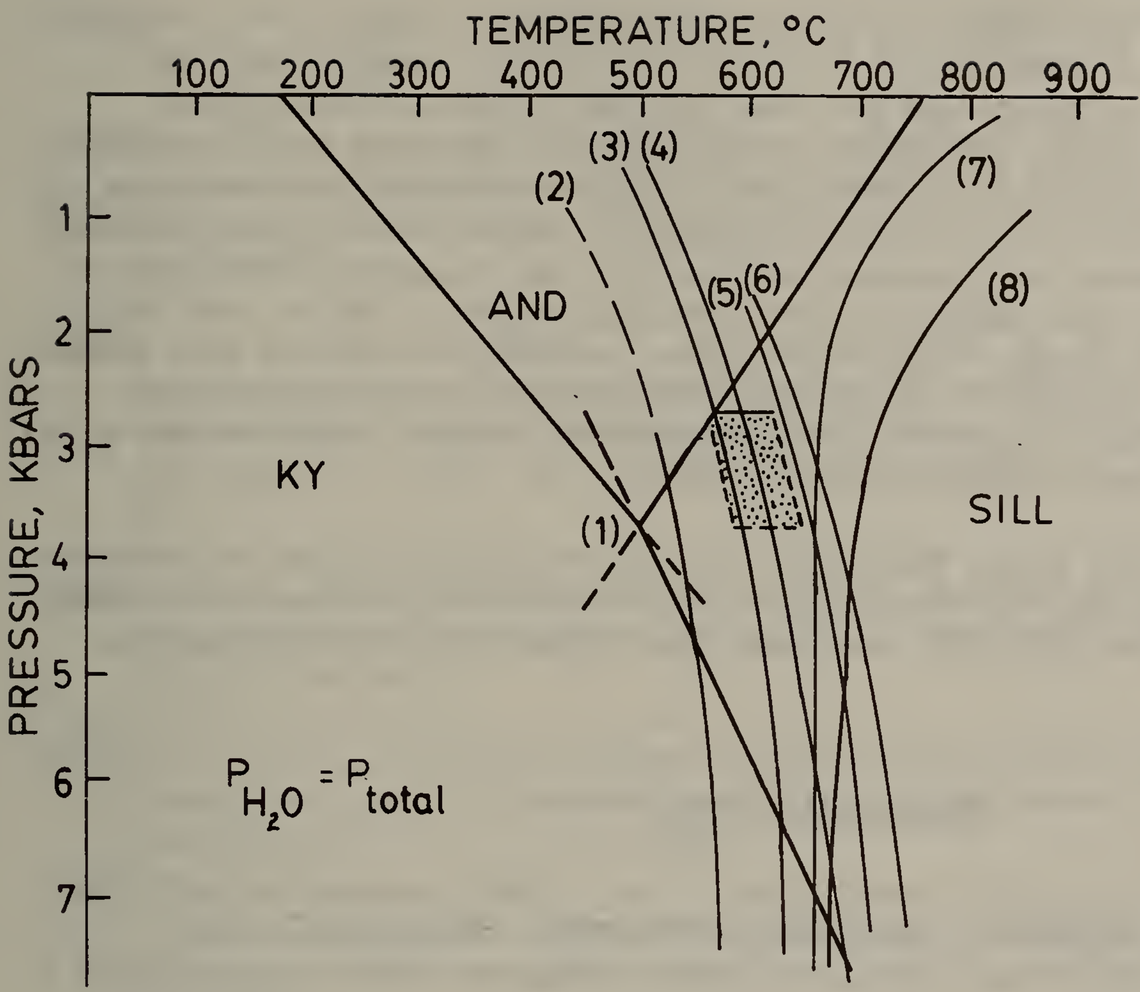
The gneisses of both the Rangeley and Littleton Formations lie within the

upper sillimanite zone. The composition of the Rangeley gneisses is believed to be very close to that of the Rangeley schists seen at Stop 2A. The assemblage Bio + Gar + Ksp (microcline) indicated on the AFM topology for the upper sillimanite zone rocks on figure 7C has been identified only in the Rangeley gneisses. However, the assemblage Sill + Ksp has not been observed in either the schists or the gneisses of the upper sillimanite zone. On an AKNa diagram for the lower and upper sillimanite zones (fig. 7D), the mineral assemblages for the Rangeley gneisses plot within the Sill + Musc + Plag field, the Musc + Plag field, and the Musc + Plag + Ksp field just as for the non-gneissose upper sillimanite zone Rangeley Formation. The tie line between muscovite and plagioclase <u>has not been broken</u>. This clearly demonstrates that the gneisses in this area have not been metamorphosed to K-feldspar + sillimanite grade. The temperature and pressure of metamorphism of the schists and gneisses can be constrained using the petrogenetic grid of figure 9. The topology for these rocks must lie to the left (lower T) of the Ksp + Sill "in" curve (curve 5 on figure 9). The presence of andalusite (formed by the reaction of curve 4 in the andalusite field) in the rocks less than a mile west of the area of Stop 2 constrains the pressure of metamorphism. The andalusite grew during M1 so the pressure for that event could not have been much higher than 2.6 kb. The metastable persistence of andalusite in the upper staurolite and lower sillimanite zones of M2 suggests that the pressure for M2 could not have been much higher than that of the triple point of Holdaway (1971). The stippled area on figure 9, so constrained, lies well to the left of the melt curves 7 and 8 suggesting that the temperatures and pressures of the M2 metamorphism were lower than those needed to allow melting. The An content of the plagioclase ranges from 18 to 22 for both the schists and gneisses of the Rangeley Formation (Billings and Fowler-Billings, 1975). This amount of Ca is not sufficient to offset the reactions described above and thus to affect the arguments advanced. The transition from schist and granulite to gneiss may simply be due to metamorphic differentiation resulting in the formation of leucosomes of quartz + plagioclase + muscovite. Further discussion of this process will take place on the outcrop at Stop 2.

STRUCTURE

All of the stratified rocks of the area are variably deformed. The structural analysis by Billings (1941) is excellent, and his maps (1941, pl. 10; Billings and others, 1946, pl. 1) clearly show the nature of folds by the outcrop pattern of the Fitch (Boott). This pattern is very similar to that shown by Hatch and Moench (1984) and on figure 3 for the Smalls Falls and Madrid. Bedding generally dips moderately to steeply and strikes predominantly to the north or northeast. Schistosity most commonly is roughly parallel to bedding. Most observed minor folds fold both bedding and schistosity and generally have axial surfaces that strike north and dip steeply. These folds are tight to open and have axes that typically plunge gently to the north or south.

In addition to the more obvious folds, which fold both bedding and early schistosity, isoclinal folds that predate them are present. These isoclines are most readily documented by reversals in topping direction of graded beds and thus are most recognized in the well-graded rocks of the lower part of the Littleton Formation. Other than a few minor folds, these structures will not be emphasized on this field trip.



(1) Al_2SiO_5 Triple Point (Holdaway, 1971)

(2) Chl + Musc ⇌ Staur + Bio + Qtz + H_2O (Hoschek, 1969)

(3) Staur + Chl + NaMusc + Qtz ⇌ Bio + Sill + KricherMusc + Ab + H_2O (see Guidotti, 1974)

(4) Staur + NaMusc + Qtz ⇌ Sill + Bio + Kricher Musc + Ab + Garn + H_2O (Guidotti, 1970)

(5) NaMusc + Plag + Qtz ⇌ Al_2SiO_5 + NaKsp + H_2O (Thompson, 1974)

(6) Musc + Qtz ⇌ Sill + Ksp + H_2O (Evans 1965)

(7) H_2O Saturated Granitic Melt (Tuttle and Bowen, 1958)

(8) H_2O Saturated Melt without K-feldspar (see Thompson, 1974)

Figure 9. P-T curves relevant to the upper staurolite through upper sillimanite zones. The stippled area indicates the interpreted approximate range of pressures and temperatures for the M2 metamorphism.

PLUTONIC ROCKS

The principal plutonic rock in the field trip area is the body of light-gray two-mica granite that underlies and extends north and northeast from the Dolly Copp Campground (body of CDg between Gorham and Stop 1 on fig. 3). Small dikes and sills of similar granite and pegmatite are common throughout the area and will be seen on the trip. This granite is similar to other Two-mica granites throughout western, central, and northern New Hampshire that have long been considered to be Devonian (Acadian) in age and assigned to the New Hampshire Plutonic Suite (New Hampshire plutonic series of Billings, 1956). Recent Rb/Sr studies of two such bodies in southern and southwestern

New Hampshire, however, gave ages of 275±10 and 330±3 Ma respectively (Lyons, 1979), and the two-mica granite of the Sebago batholith (eastern edge of figure 3) has recently given an age of about 325 Ma by both U-Pb zircon (Aleinikoff and others, 1985) and Rb/Sr determination (Hayward and Gaudette 1984). Arth and Ayuso (1985) have confirmed by Rb-Sr studies the Devonian age of similar two-mica granites in northeast Vermont. Until additional isotopic dating is done on the two-mica granites of eastern New Hampshire, the age of these bodies and of the high-grade regional metamorphism to which they seem to be closely related is in doubt. The high-grade rocks of the field trip area are only about nine miles west of exposures of pink two-mica granite and pegmatite identical to those dated by Aleinikoff and others (1985) a few miles further east in the Sebago batholith (fig. 3).

Plutonic rocks of the White Mountain Plutonic-Volcanic Suite are abundant in eastern New Hampshire north, west, and south of the field trip area. They are the subject of another trip at this meeting (Trip B-2) and will not be covered on this trip.

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ROAD LOG FOR TRIP B-3

Pertinent maps:
- Topographic maps:
  - Carter Dome, N. H. 7.5-minute (1:24,000)
  - Mt. Washington, N. H. 7.5-minute x 15-minute (1:25,000)
  - or Mt. Washington, N. H. 15-minute (1:62,500)
  - Crawford Notch, N. H. 15-minute (1:62,500)
  - North Conway, N. H. 15-minute (1:62,500)
- Geologic maps:
  - Gorham 15-minute (Billings and Fowler-Billings, 1975)
  - Crawford Notch 15-minute (Henderson and others, 1977)
  - Mt. Washington 15-minute (Billings and others, 1946)
  - North Conway 15-minute (Billings, 1928)
  - White Mtn. region (Hatch and Moench, 1984)
  - New Hampshire (Billings, 1956)

Trip will assemble at the south end of the Dolly Copp Campground west of Route 16, south of Gorham. To reach the assembly point turn west off N.H. Route 16 onto the Dolly Copp Road (paved) about 4.3 miles south of Gorham, N.H., or about 2.5 miles north of the entrance to the Mount Washington Auto Road. Turn at U.S.F.S. sign for Dolly Copp National Forest Campground. After 0.3 mile turn left (south) into the entrance to the Campground. Proceed south on paved road through the campground about 0.9 mile to the assembly point at the south end of the campground.

From assembly point walk south along the Great Gulf Link Trail (marked simply Great Gulf Trail on both the Gorham 15-minute and the Carter Dome 7.5-minute quadrangles) about 1700 feet (520 meters) to outcrops in the Peabody River immediately east (left) of the trail (fig. 10).

STOP 1 Rangeley, Smalls Falls, Madrid, and Littleton Formations (fig. 10). This first exposure in the river is Rangeley Formation. Rocks are well-bedded, gray, slightly rusty weathering mica schist and granulite in which schist beds are 2-10 cm thick and granulite beds are 1-3 cm thick (fig. 5). Some beds are graded and consistently indicate tops are to the west. Small knots of muscovite are interpreted to be pseudomorphs after both staurolite and andalusite. Lenticular pods ("footballs") of calcareous granulite 15-30 cm thick and 0.5-1 meter long are common (fig. 5). Whether they represent concretions or boudinaged beds is a question that can be debated.

From this exposure of Rangeley in the river proceed west through the woods up the slope of hill 2584 (fig. 10). At about elevation 1600 feet, after passing over exposures of well-bedded Rangeley Formation, we will look at an exposure of metamorphosed grit in which the individual clasts are 2-4 mm across (some are slightly larger). Channels of this grit into underlying schist are locally exposed (fig. 11), and all indicate that the section youngs to the west. This grit is shown on Billings and Fowler-Billings' (1975) map as a horizon of quartz conglomerate. We suggest that it supports our correlation of this group of rocks with the Rangeley Formation.

At about elevation 1670 feet in a small gulley are exposures of about 25 meters of section of deeply rusted, slabby, flaggy sulfidic gray quartzite and minor schist. These rocks are typical of the Smalls Falls Formation, to which we assign them.

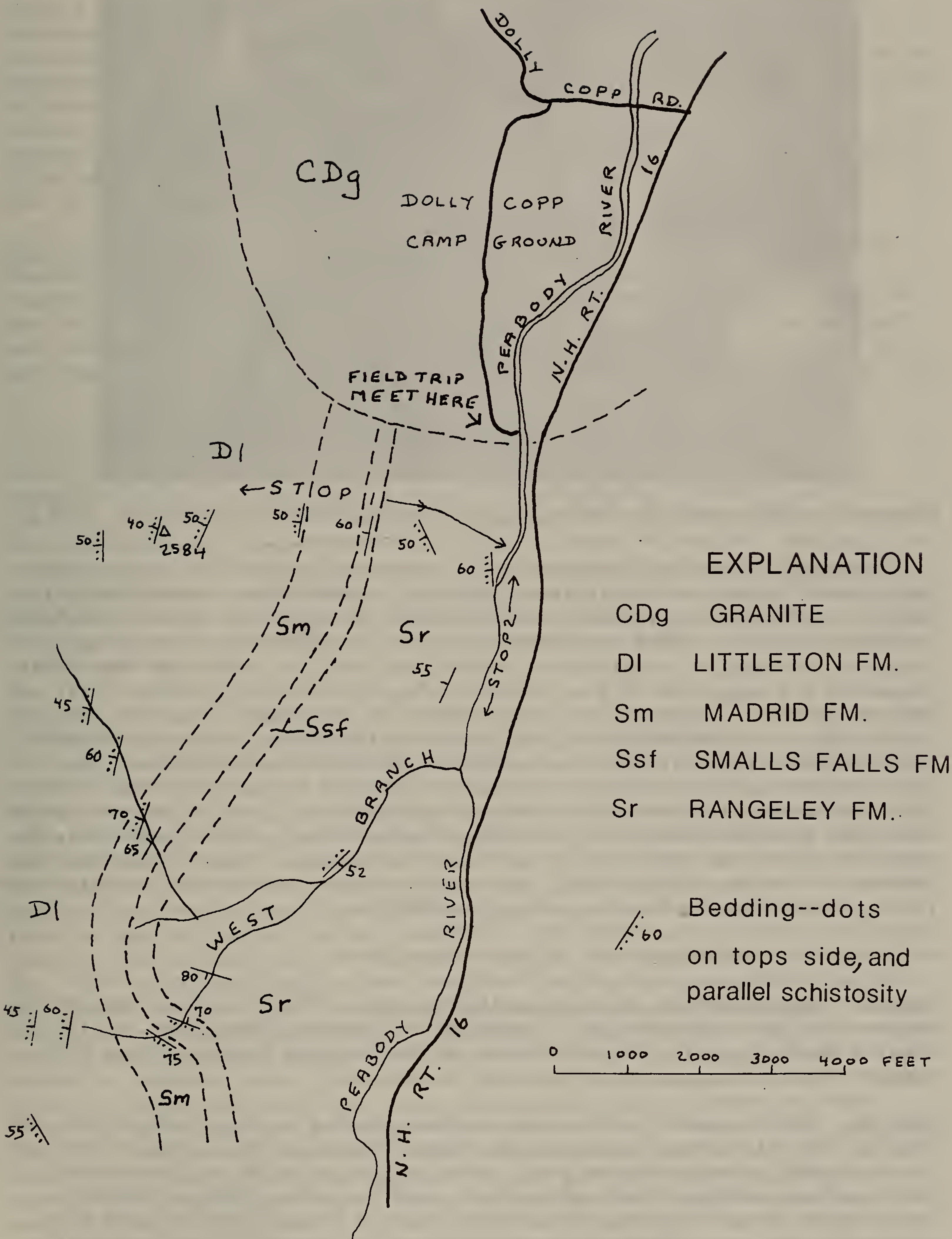

Figure 10. Geologic map of the rocks in the vicinity of Stops 1 and 2.

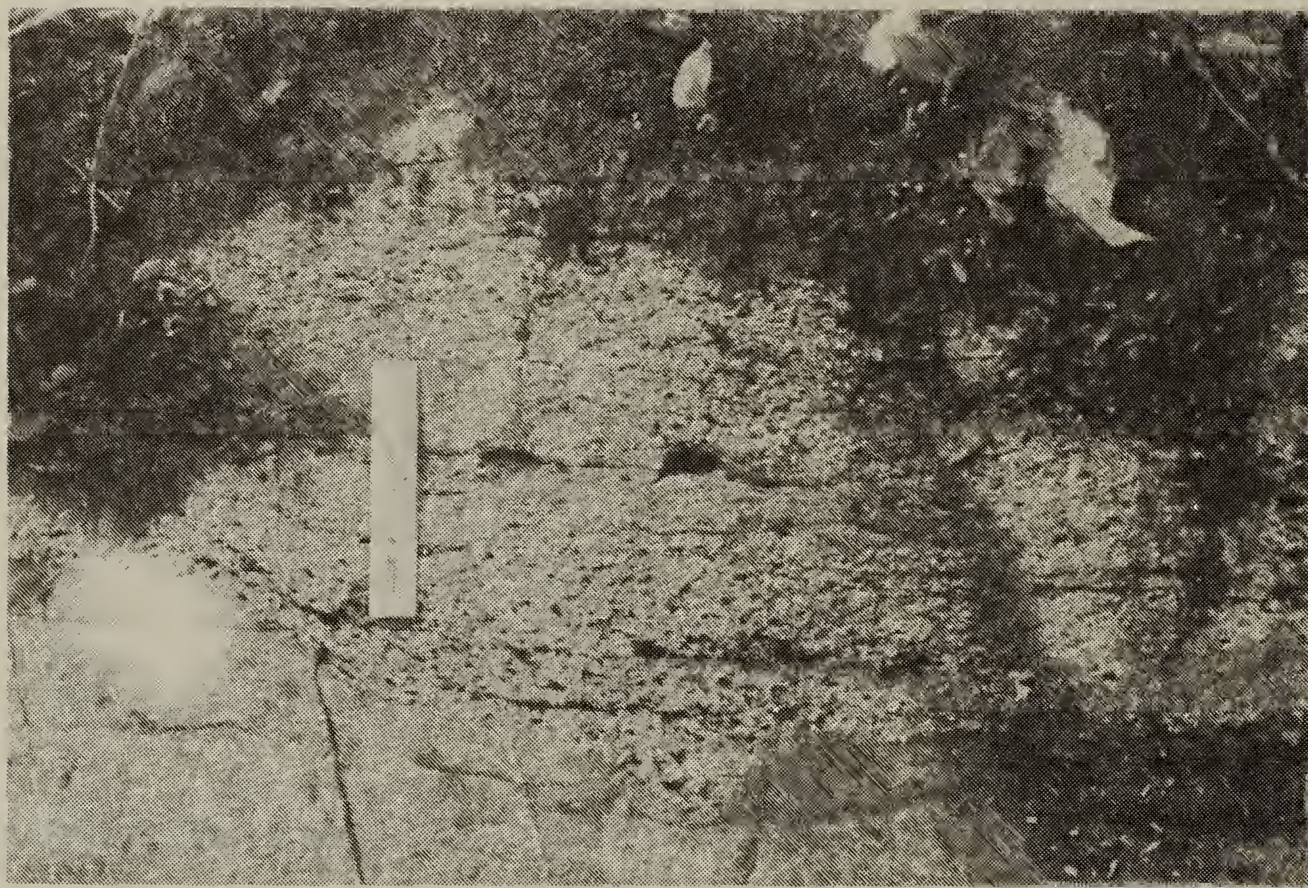

Figure 11. Photograph of channel of grit in the Rangeley Formation. Stop 1.

At about elevation 1740 feet, about 70 meters south of the gulley, are exposures of green and light-green, nonrusty, nonsulfidic, compositionally layered calc-silicate granulite characteristic of the lower part of the Madrid Formation (fig. 4). This rock typically consists primarily of actinolite, diopside, microcline, plagioclase, quartz, and biotite. Upslope from these exposures are exposures of gray nonrusty plagioclase-quartz-biotite "salt and pepper" granulite, typical of the upper part of the Madrid Formation, and upslope again are exposures of gray nonrusty sillimanite schist interbedded with micaceous quartzite that we assign to the Littleton Formation. The Madrid-Littleton contact appears to be gradational, and we have drawn it at the point where aluminous schist becomes a significant part of the rock going upsection from the thick "salt and pepper" granulites. Local graded beds in the Littleton rocks consistently indicate stratigraphic tops to the west all the way to the top of hill 2584. The trip will turn around and head back down to the starting point by about elevation 2000 feet.

Return to starting point in the Peabody River.

<u>Stop 2</u>. <u>Transformation of the Rangeley Formation</u>. This stop will include a sequence of four exposures going south up the Peabody River (figs. 6, 10) to observe the progressive transformation of the Rangeley Formation from well-bedded schist and granulite to gneiss.

<u>Stop 2A</u>. This exposure is the same as that examined as the first outcrop in Stop 1. On this return visit we will stress the metamorphic, rather than the stratigraphic, aspects of the rock. Note particularly the well-preserved nature of the bedding, local graded beds, and the calc-silicate footballs, even though these rocks are sufficiently high in the sillimanite zone that staurolite has gone. Abundant pseudomorphs after staurolite and andalusite range from 1 to 4 cm in length and are composed of muscovite and lesser

amounts of quartz, plagioclase, biotite, and sillimanite. Sillimanite crystals 0.1-0.8 mm long are common in the pseudomorphs, whereas in the groundmass mats of fibrolite have nucleated on biotite grains. Aggregates of prismatic sillimanite are also present in the groundmass. A few small lenses of two-mica granite and pegmatite are present. Walk upstream along the outcrop about 200 feet to a point where the stream forks (actually this is the downstream end of a long thin island in the main body of the Peabody River). Climb up onto the west bank of the river and continue upstream for about 250 feet to

Stop 2B. (fig. 6) Outcrop in the Peabody River immediately beside the now abandoned trail on the west side of the river. Rock is slightly rusty-weathering sillimanite schist and granulite. Bedding is well preserved, as are calc-silicate "footballs". Two-mica granite and pegmatite are slightly more abundant than at Stop 2A. Continue south upstream along old trail on west bank of river about 1350 feet south of Stop 2A to next exposures in river at

Stop 2C. (fig. 6) Rock is again sillimanite rich and slightly rusty weathering, but here it is relatively massive and, although lenses of granulite and schist are locally present, throughgoing beds are rare to absent. Calc-silicate "footballs" are common, as are distinct spangles or clots of muscovite coarser than those seen at Stops 2A and 2B. Thus in this exposure we see the beginning of degradation of bedding and the beginning of gneissic character and banding. Note the contrast between this rock and the boulders of gray, nonrusty Littleton Formation schist in the river. Continue south upstream another 400 feet to large exposures in the river at the foot bridge over the river.

Stop 2D. (fig. 6) This rock is a moderately rusty gneiss. Much of the rust has the brick-red color characteristic of the Rangeley Formation. Calc-silicate "footballs" are abundant. Bedding is nowhere recognizable in the outcrop, but the trend of bedding and foliation in all of the previous exposures seen in Stop 2 suggest that we have been traversing approximately along the same stratigraphic horizon. Blocks of gray plagioclase-quartz-biotite granulite are preserved within the gneiss. Irregular blobs of quartz-feldspar-muscovite-biotite-garnet-tourmaline pegmatite are common throughout the outcrop, and the small lenticular aggregates of quartz and feldspar that form the gneissic foliation are parallel to the schistosity in the schistose lenses. This schistosity, in turn, is parallel to the bedding at Stops 2A and 2B. These relations suggest that incipient melting has segregated quartzo-feldspathic material parallel to the bedding and schistosity of the bedded Rangeley to produce the gneissic foliation. Irregular bodies of gneiss within some of the pegmatites suggest inclusion of gneiss, but some of these gneiss-pegmatite contacts appear gradational, and some of the pegmatites are slightly foliated, suggesting that they were present before the deformation ended. Were the pegmatites, and associated two-mica granites, derived very locally by incipient anatexis or were they formed elsewhere and moved into their present site? And finally, were the metamorphism and granite/pegmatite formation Acadian or Alleghanian events? When we have resolved these interesting questions, we will return to the cars via the Great Gulf Link Trail.

From the original assembly point (mileage 0.0) drive back north out of the campground.

Mileage

0.9 Turn right (east) onto Dolly Copp Road.

1.3 Turn right (south) onto Route 16.

2.1 Entrance to Dolly Copp Picnic Area on right. Continue south on Route 16.

2.9 Outcrop on left (east) side of road at Greens Grant/Martins Location town line is slightly rusty muscovite-spangled gneiss assigned to the Rangeley Formation.

3.8 Small outcrops on right (west) and in river below it are moderately rusty gneiss with calc-silicate "footballs" assigned to the Rangeley, plus pegmatite and binary granite.

4.7 Entrance to Mount Washington Auto Road on right. Continue south on Route 16.

4.8 Small outcrops on left of nonrusty gneiss assigned to the Littleton Formation.

5.1 Enter Pinkhams Grant.

5.2 Outcrop on left of slightly rusty to nonrusty gray gneiss assigned to the Littleton Formation, with much pegmatite and two-mica granite.

5.4 Pull into paved parking on right and park. From a point 130 feet north of the south end of the parking area climb down to the Peabody River for

Stop 3A. Smalls Falls, Madrid, and bounding gneisses at Emerald Pool. The exposures to be examined are on both sides of a small beach about 20 feet wide on the east side of the river. At the south edge of the beach is about 10-15 feet of deeply rusty, sulfidic, flaggy quartzite and schist characteristic of and assigned to the Smalls Falls Formation. South of this rusty quartzite and schist is decreasingly rusty weathering gneiss containing calc-silicate "footballs". At the north side of the beach is about 10 feet of nonrusty, well-banded, light- and dark-green calc-silicate granulite characteristic of the basal part of the Madrid Formation, bounded on the north by nonrusty, gray, muscovite-rich gneiss. The difference between the rusty "football"-bearing gneiss to the south and the nonrusty nonfootball-bearing gneiss to the north is very obvious here. We conclude that the rusty "football"-bearing gneiss is Rangeley and that the nonrusty gneiss is either upper Madrid or lower Littleton. On the basis of the high muscovite content we lean toward lower Littleton. If these interpretations are correct, both the Smalls Falls and the Madrid are less than 20 feet thick at this locality, in contrast to the much thicker sections of both formations seen at Stop 1. Does this mean that Stop 3 was originally closer to the shore (more proximal) than Stop 1 even though Stop 3 appears now to be slightly further east of the axis of the Bronson Hill anticlinorium than Stop 1? Or do the differences in present

thickness simply reflect differential tectonic thinning? Of interest also is the observation that although both the Littleton rocks and the Rangeley rocks at this locality have been thoroughly converted to gneiss, the Smalls Falls sulfidic quartzites and the Madrid calc-granulites which must have gone through the same metamorphic conditions have resisted the "gneissification" process. Other examples of this same phenomenon were observed throughout a large area of central eastern New Hampshire and adjacent westernmost Maine.

Climb back up to the cars and cross over Route 16 to the cut on the east side of the highway for

Stop 3B. Projection of the exposure of Smalls Falls and Madrid at Emerald Pool says that this road cut of gneiss is north of them and thus that the cut is in upper Madrid or Littleton. Although some local rusting is visible in the cut, it is not as rusty weathering as most exposures of Rangeley gneiss. Furthermore, the fresh rock is gray, not brown or gray-brown, and calc-silicate "footballs", if present, are extremely rare in the cut. The gneissic banding is highly contorted and swirled, and the gneiss is laced with tourmaline-rich pegmatite and two-mica granite. No trace of original bedding can be detected. Clots of muscovite 1-5 cm across are believed to be pseudomorphous after andalusite. The high percentage of muscovite suggests that the protolith of the gneiss was aluminous schist, and thus we interpret this gneiss to be Littleton Formation rather than upper Madrid.

Return to cars and continue south on Route 16.

5.8 Pull into paved parking area on the right (west) side of highway and walk down onto the outcrops in the river from the north end of the parking area.

Stop 4. Rangeley Formation (?) gneiss. Rock here is moderately rusty-weathering gneissic granite or granitic gneiss with abundant "footballs", some as much as a meter across, that are compositionally layered suggesting original bedding. The "footballs" and their internal layering are oriented in all directions, and some contain folds in their layering. Regional stratigraphic and structural relations say that this rock should be Rangeley Formation, and both the presence of rusting and the calc-silicate "footballs" support this conclusion. The degree of gneissification here is extreme, however, and locally the rock looks like a slightly foliated granite. Does this rock represent introduced granitic melt, with inclusions of Rangeley calc-silicate "footballs", which was subsequently somewhat deformed to produce the foliation, or does it represent intensely metamorphosed and migmatized Rangeley Formation?

Return to the cars and continue south on Route 16.

6.6 Wildcat Mountain Ski area on left.

6.7 Outcrop on right of moderately rusty-weathering two-mica granite and pegmatite.

7.5 Turn right at sign "Pinkham Notch Camp, Appalachian Mountain Club" into parking area and park. From steps up from parking area walk about 1600 feet (500 meters) up

the Tuckerman Ravine Trail to wooden bridge over the Cutler River. [300 feet further up trail is outlook for Crystal Cascade--a very pretty waterfall in the Cutler River.] From the wooden bridge go up the river over ledges of two-mica granite and pegmatite and gray post-metamorphic dike rock presumably of the White Mountain Plutonic-Volcanic Suite.

Stop 5A. Smalls Falls and Madrid Formations. 30-45 meters above the wooden bridge on both sides of the river are outcrops of rusty-weathering, flaggy, sulfidic schist and quartzite bounded on the west (upstream) by nonrusty-weathering green calc-silicate granulite. We map these rocks as the top of the Smalls Falls and the basal beds of the Madrid Formation. Beds and parallel schistosity here strike about north-south and dip about 40° west.

Return to the bridge and walk down the river across outcrops of two-mica granite and pegmatite for about 250 feet to

Stop 5B. Smalls Falls and Rangeley Formations. At this point cross contact from two-mica granite into deeply rusty, thinly bedded (1-3 cm), dark-gray, graphitic schist and quartzite containing much pyrrhotite in the fresh rock. We map this rock as Smalls Falls. Stream flows roughly parallel to beds and schistosity in this rock for about 30 meters, then swings southeast across bedding, exposing the contact between the rusty sulfidic Smalls Falls to the west and only slightly rusty well-bedded schist and granulite to the east. This latter rock contains lenticular "footballs" of calc-silicate granulite, and we map it as Rangeley Formation. The contact is exposed and can be pinpointed to an interval of a few inches. About 10 meters of well-bedded Rangeley schist and granulite are exposed before the outcrop runs out downstream. Note that in contrast to the gneiss seen at Stops 2D and 4, the Rangeley rock exposed here retains good bedding. Grading is questionable, but a few beds in both the Rangeley and the Smalls Falls may indicate tops to the west. Across the valley to the east, all of the rocks are intensely gneissic.

Walk about 10 meters north from the river to the Tuckerman Ravine Trail and walk back down the trail to the cars.

This is the end of the field trip.

For the shortest route back to Lewiston, turn left (north) from the Pinkham Notch Camp parking lot onto Route 16 and follow Route 16 north to Gorham, N. H. In Gorham turn east onto Route 2 to Bethel, Maine, and take Route 26 southeast through Norway. At Welchville turn left onto Route 121, which joins Route 202 in Auburn. Cross the Androscoggin into Lewiston and Bates College.

STRATIGRAPHIC AND STRUCTURAL RELATIONSHIPS BETWEEN THE CUSHING, CAPE ELIZABETH, BUCKSPORT, AND CROSS RIVER FORMATIONS, PORTLAND-BOOTHBAY AREA, MAINE.

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INTRODUCTION

The purpose of this field trip is to examine the general geologic relations between the Cape Elizabeth, Cushing, Bucksport, and Cross River Formations between Portland and Boothbay, Maine. The general geology of the region is covered in the description for field trip A4 and the reader is referred to that section for geological background for this, the B4, trip. Figure 1 is a generalized geologic map of southwestern Maine, and Figure 2 shows the itinerary and general location of stops for this trip.

ITINERARY

The assembly point is in front of Chase Hall on the Bates Campus, and the trip is by 15 passenger vans. The assembly time is 8:00 A. M. The following road log begins at the end of the Maine Turnpike at the commuter parking lot at exit 7 in South Portland.

Mileage	
0.0	At commuter parking lot for exit 7, Maine Turnpike South Portland. Exit right onto turnpike connector.
1.7	Junction with Route 1; turn left.
2.4	Bridge over Boston and Maine tracks.
2.7	Right on Broadway at stoplight.
3.3	Stoplight. Continue straight on Broadway.
3.5	Stoplight. Bear left staying on Broadway
4.9	Stoplight at Ocean Avenue. Continue straight on Broadway.
5.1	Stoplight. Turn right on Cottage Road.

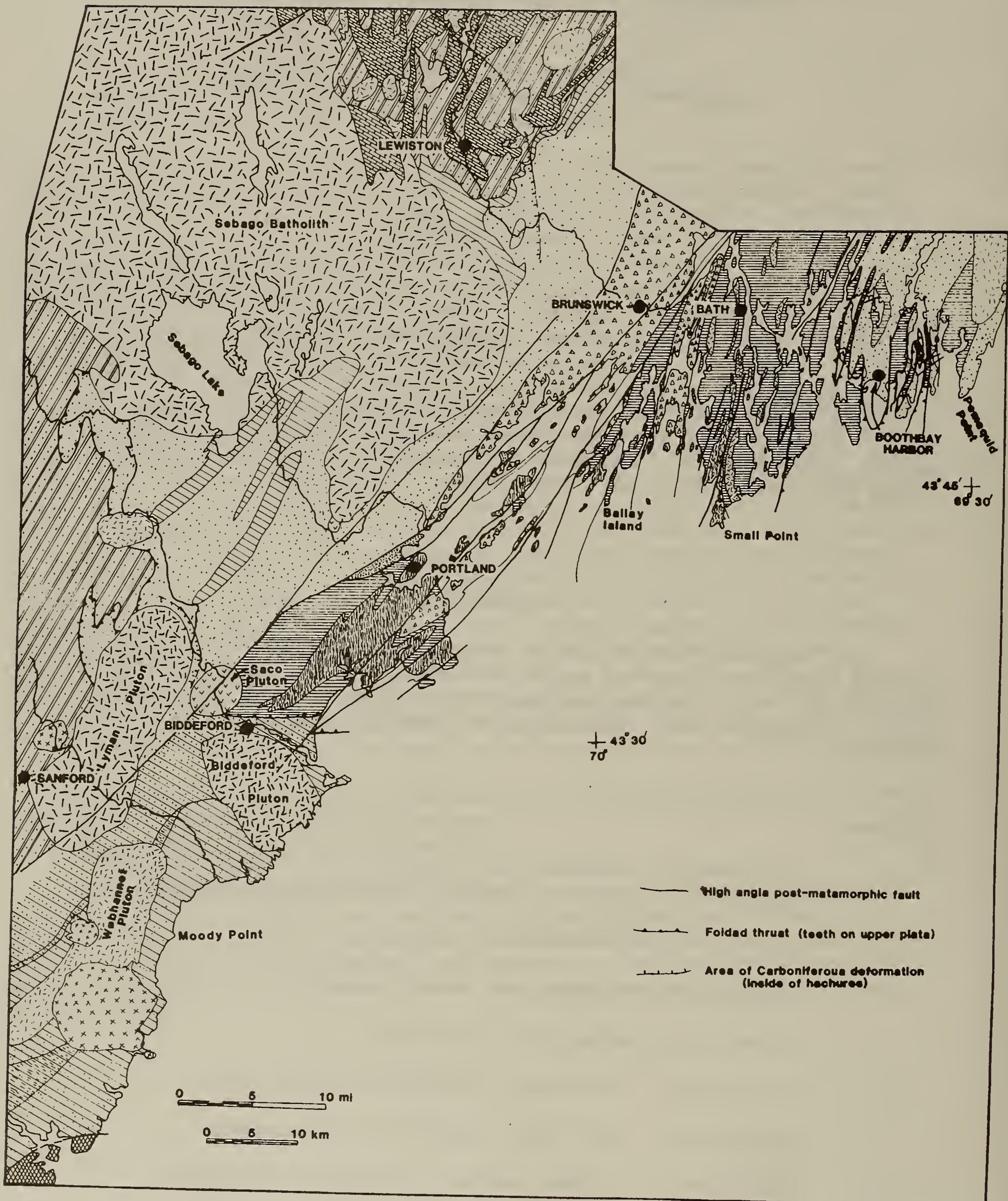


Figure 1. General geology of southwestern Maine

EXPLANATION

INTRUSIVE ROCKS

Mesozoic	Gabbro, alkaline granite,and related intrusives
Carboniferous	2-mica granite
Carboniferous or older	Foliated gabbro-diorite
E. Devonian	2-mica granite
	Foliated granodiorite

STRATIFIED ROCKS

E. Sil.	Rindgemere, Sangerville Fms, Patch Mtn. M.,Sangerville	CENTRAL MAINE SEQUENCE
	Windham, Waterville Fms., Anasagunticook M.,Sangerville Fm.	
L. Ord. to E. Sil.	Vassalboro Fm.	
Pre€ ?	Berwick Fm.	MERRIMACK GROUP
	Eliot Fm.	
	Kittery Fm.	
Pre€ ? to Ord ?	Macworth Fm.	CASCO BAY GROUP
	Jewell,Spurwink,Scarboro, Diamond Island, Spring Point Fms.	
	Cape Elizabeth Fm.	
	Cushing Fm.	
Pre€	Rye Fm.	

L. Ord. to E. Dev.

Bucksport Fm.

Cross River Fm.

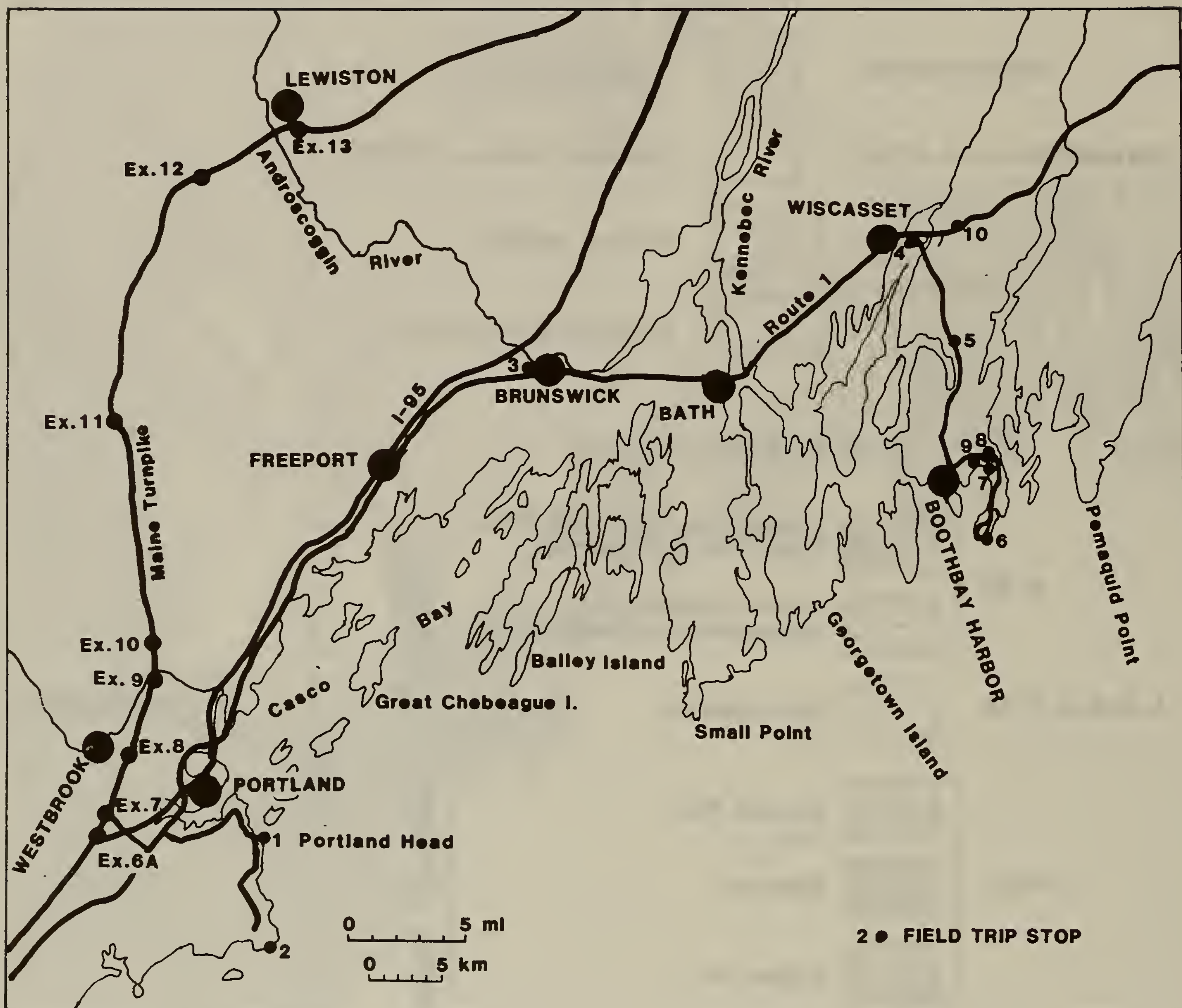


Figure 2. Itinerary of trip B4.

6.3 Junction with Prebble Street. Cottage Road now becomes Shore Road. Continue on Shore Road.

7.1 Turn left into Fort Williams Park. Follow road to:

7.7 Portland Headlight parking area adjacent to the lighthouse.

STOP 1 (Figure 3).

Figure 3. Location of Stop 1. Cape Elizabeth and Portland E. 7.5′ quadrangles.

Walk along path to the north of the lighthouse until you can easily descend down over the seacliff. The Cushing Formation exposed here, close to the crest of the Cushing Anticline, is light gray quartz + plagioclase + biotite gneiss with minor muscovite and microcline. Pyroclastic structures (relict crystal fragments and breccia clasts) are rare here suggesting that these rocks represent original fine-grained felsic tuffs. Approximately 100 meters north of the lighthouse, close to the base of the seacliff, light gray metatuff can be seen in contact with, and overlying a thin zone of buff-colored metatuff probably representing two separate ash deposits. The lack of interbedded volcanogenic metasediments and the lack of

well-developed compositional layering indicative of current reworking or sorting of pyroclastic debris, are suggestive of a subaerial origin of this part of the Cushing. On the other hand, the presence of manganiferous metasediments (Wilson Cove Member of the Cushing) in contact with the top of the Cushing nearby on the east side of the Cushing Anticline suggests a subaqueous origin for this metafelsite.

Structurally, the metafelsite is characterized by a weak foliation and a strong lineation plunging gently to the southwest, parallel to axes of F2 structures. Several unmetamorphosed basalt or diabase dikes of presumed Juro-Triassic age cut the Cushing in the general vicinity of the lighthouse. About 100m north of the lighthouse a hackly-fractured postmetamorphic felsic dike occupies a high-angle brittele fault zone, and has been internally faulted parallel to its contacts with the Cushing.

At the base of the seacliff immediately below the northeast side of the lighthouse is a rare occurrence of secondary copper minerals and limonite, apparently the result of interaction of seawater with a large batch of discarded brass screws, wire, and other hardware items!

Return to vehicles and retrace route to entrance to Fort Williams Park.

8.3 Turn left onto Shore Road at Fort Williams Park entrance.

9.6 Blinker. Turn left onto Route 77.

10.7 Roadcuts of rusty-weathering Scarboro Formation.

10.9 Roadcut of Spurwink Metalimestone in the axis of the Peables Point syncline, a refolded recumbent syncline.

11.3 Left turn onto road to Two Lights State Park.

11.4 Left turn onto Two Lights Road

12.4 Turn right into entrance to Two Lights State Park and proceed to parking lot at end of the road.

STOP 2 (Figure 4).

Walk to shoreline exposures. Since this is a state park, no collecting is permitted, and no hammers should be used.

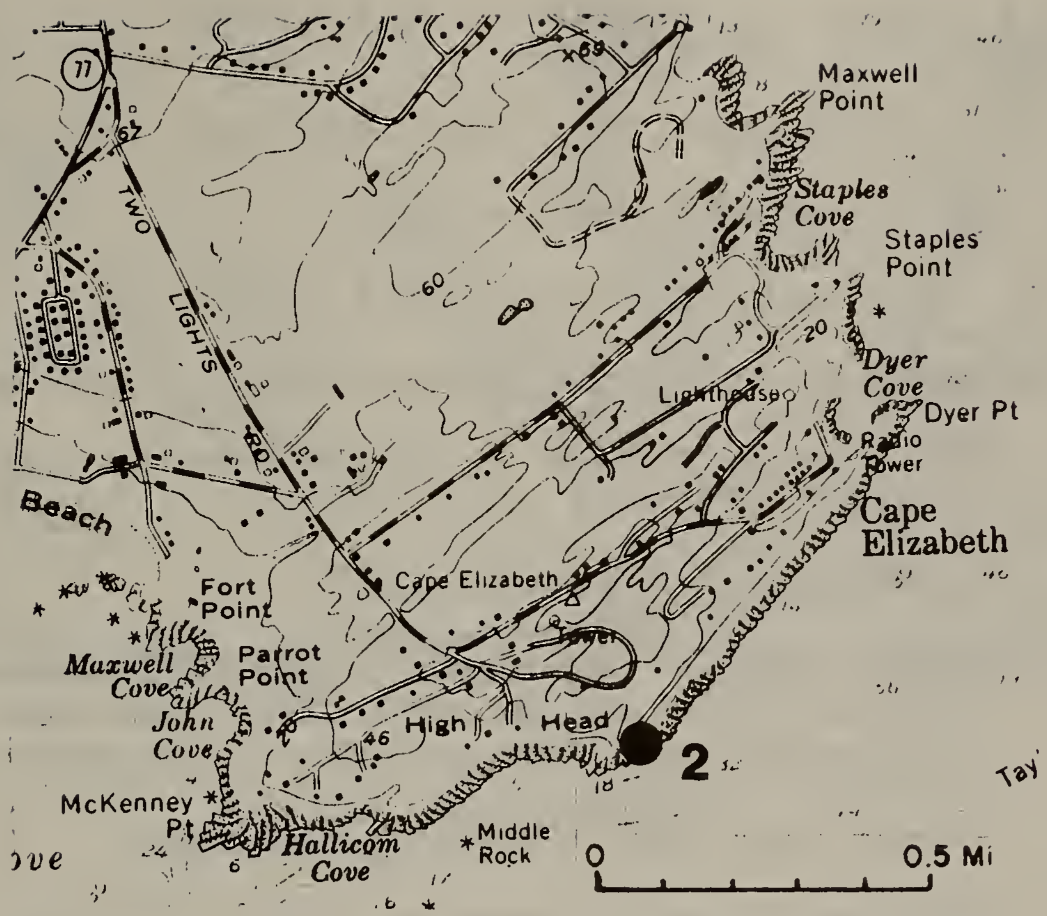


Figure 4. Location of stop 2 (Cape Elizabeth 7.5′ quadrangle).

The exposures here are chlorite-grade Cape Elizabeth Formation consisting of thin- to thick-bedded, buff-weathering metasiltstone and dark gray phyllite. The metasiltstone beds are moderately calcareous and ankeritic, hence their buff-weathering color. The exposures here are a bit atypical of the Cape Elizabeth Formation at similar grade elsewhere in that the metasiltstone beds are much thicker, occasionally exceeding 1 meter. This belt of low grade Cape Elizabeth is separated from biotite to garnet grade rocks of the Spring Point Formation which abuts it approximately 1 km to the northwest, by the post-metamorphic high angle Broad Cove Fault.

The Cape Elizabeth Formation here is affected by two major deformations. F1 folds produced by the first are east-verging recumbent folds. Upright graded bedding indicates that these folds are east-facing. F2 folds are very gentle and open. Recumbent parasitic folds have an axial planar spaced cleavage in the metasiltstone beds, and phyllitic cleavage in the dark phyllite beds. Both types of cleavage are parallel to axial planes of the F1 folds.

Return to vehicles, and retrace route to the entrance to the State Park.

13.1 Turn leftish from the State Park onto Two Lights Road.

14.0 Turn left on Wheeler Road

14.1 Stopsign. Turn left onto Route 77.

15.0 Pass entrance to Crescent Beach State Park.

16.2 Bear left at fork (Spurwink Road to the right), staying on Route 77.

16.5 Cross the Spurwink River.

16.7 Sawyer St. to the right. Stay on Route 77. The type locality for the Spurwink Metalimestone is across the marsh from here, on the banks of the tidal river.

18.0 Turn right onto Pleasant Hill Road.

19.5 Stop sign and blinker at Highland Road. Continue straight on Pleasant Hill Road.

21.4 Stoplight. Turn right onto Route 1.

21.6 Keep straight at stoplight.

22.1 Turn left at stoplight onto Turnpike connector.

22.6 Bear right on I-295 ramp.

23.3 Merge with I-295. Stay on this highway (which merges in Falmouth with I-95) to Brunswick. Roadcuts along I-95 and I-295 are all of the Vassalboro Formation, in places very extensively migmatized.

32.2 I-295 merges with I-95.

51.2 Exit right for Brunswick to Route 1.

52.2 Stoplight. Continue straight on Route 1

52.5 Turn left on Route 1.

52.8 Turn left into parking area at the end of the Topsham-Brunswick footbridge.

STOP 3 (Figure 5).

Walk across footbridge to the Topsham side of the Androscoggin River. At the end of the bridge, follow a short footpath to ledges at the waters edge. These

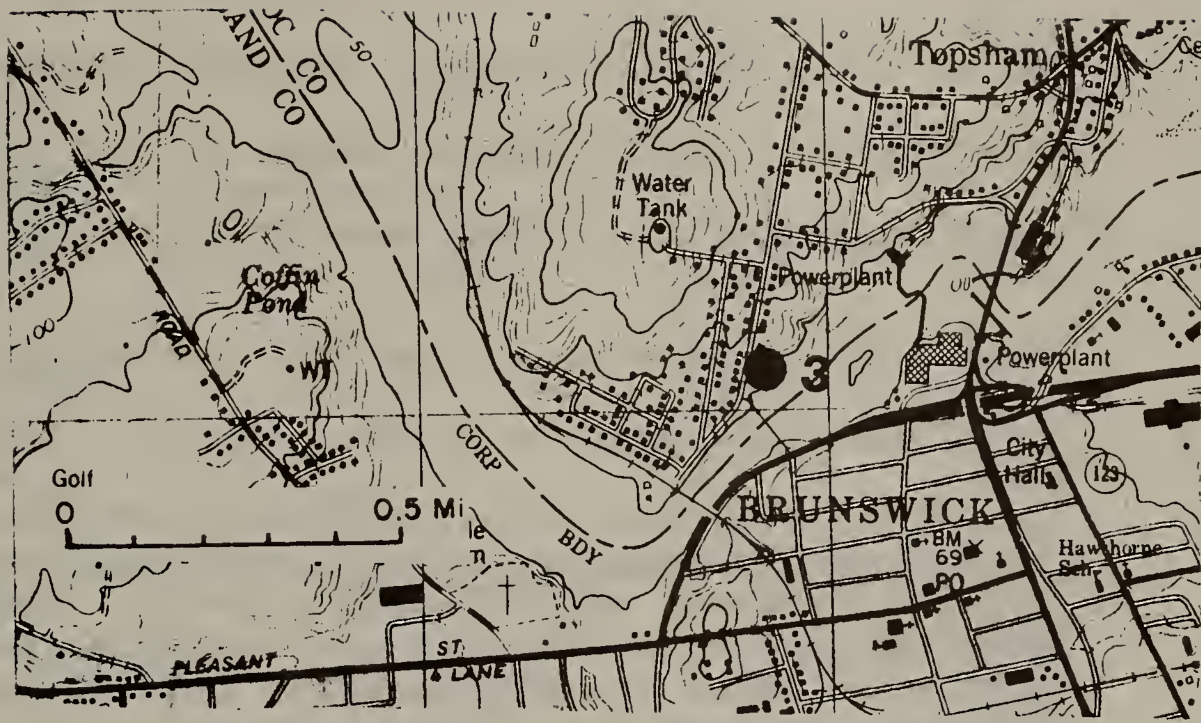


Figure 5. Location of stop 3 (Brunswick 7.5′ quadrangle).

exposures are typical of the principal lithology of the Mount Ararat Member of the Cushing Formation, here consisting of interbedded granofelsic amphibolite and light gray biotite+hornblende granofels and gneiss cut irregularly by granitic orthogneiss, protomylonitic foliated pegmatite stringers, unfoliated pegmatite dikes, and granite pods.

On the Brunswick side of the footbridge the contact with the Nehumkeag Pond Member of the Cushing may be exposed, lying essentially at the waters edge. Just upstream and slightly west of strike are interbedded biotite granofels and amphibolite typical of the Mount Ararat lithology we observed on the Topsham side of the river. The Newhumkeag Pond Member as defined by Newberg (1981) in the Wiscasset - Gardiner area to the north is a buff-weathering fine-grained quartz + K-feldspar + plagioclase + muscovite + biotite gneiss, differing from the Mount Ararat Member principally in the virtual absence of amphibolite. The same contact, apparently conformable, was exposed during construction of the hydropower dam just downstream from this stop.

Return to vehicles. Continue on Route 1 for Bath, being careful to take the underpass and not exit to Maine Street, Brunswick.

52.9 Beginning of divided Route 1 at underpass.

57.3 We are crossing the core of the Hen Cove Anticline.

58.0 New Meadows exit. Stay on Route 1.

60.2 Stay on Route 1. Do not take the exit for downtown Bath.

60.6 West end of Carleton Bridge over the Kennebec River.

61.3 Stoplight just east of bridge. Stay on Route 1.

61.6 Junction Route 127. Stay on Route 1. From here on to Wiscasset, roadcuts expose migmatized sillimanite + K-feldspar grade Cape Elizabeth. Pegmatites that cut the metasediments are characterized by pink K-feldspar, an abundance of sillimanite, and sporadic occurrences of dumortierite, chalcopyrite and bornite. Secondary copper staining is occasionally observed along freshly exposed fracture surfaces.

67.0 Junction with Route 144. Stay on Route 1.

70.4 Junction with Route 27. Stay on Route 1.

70.8 You are in downtown Wiscasset.

70.9 West end of the bridge over the Sheepscot River. Just downstream on the Wiscasset shore are the remains of the old cargo schooners Hesper and Luther Little.

71.5 East end of Sheepscot River bridge. Turn right on road immediately after the Muddy Rudder Restaurant.

71.8 Turn right onto Fort Road, following signs to Fort Edgecomb.

72.1 Turn left into Fort Edgecomb grounds and park.

STOP 4 (Figure 6). LUNCH

Inasmuch as this is a state park, collecting is not permitted, and hammering is discouraged. Walk to exposures south southeast of the block house along the shore. This is an exposure of the typical Cape Elizabeth Formation not much migmatized and here consisting of thinly bedded quartz + plagioclase + biotite + muscovite schist and granofels, with lesser metapelite. Note the abundant sillimanite and cordierite adjacent to the pegmatite stringers. Small grains of bright purplish blue dumortierite are present in the pegmatite.

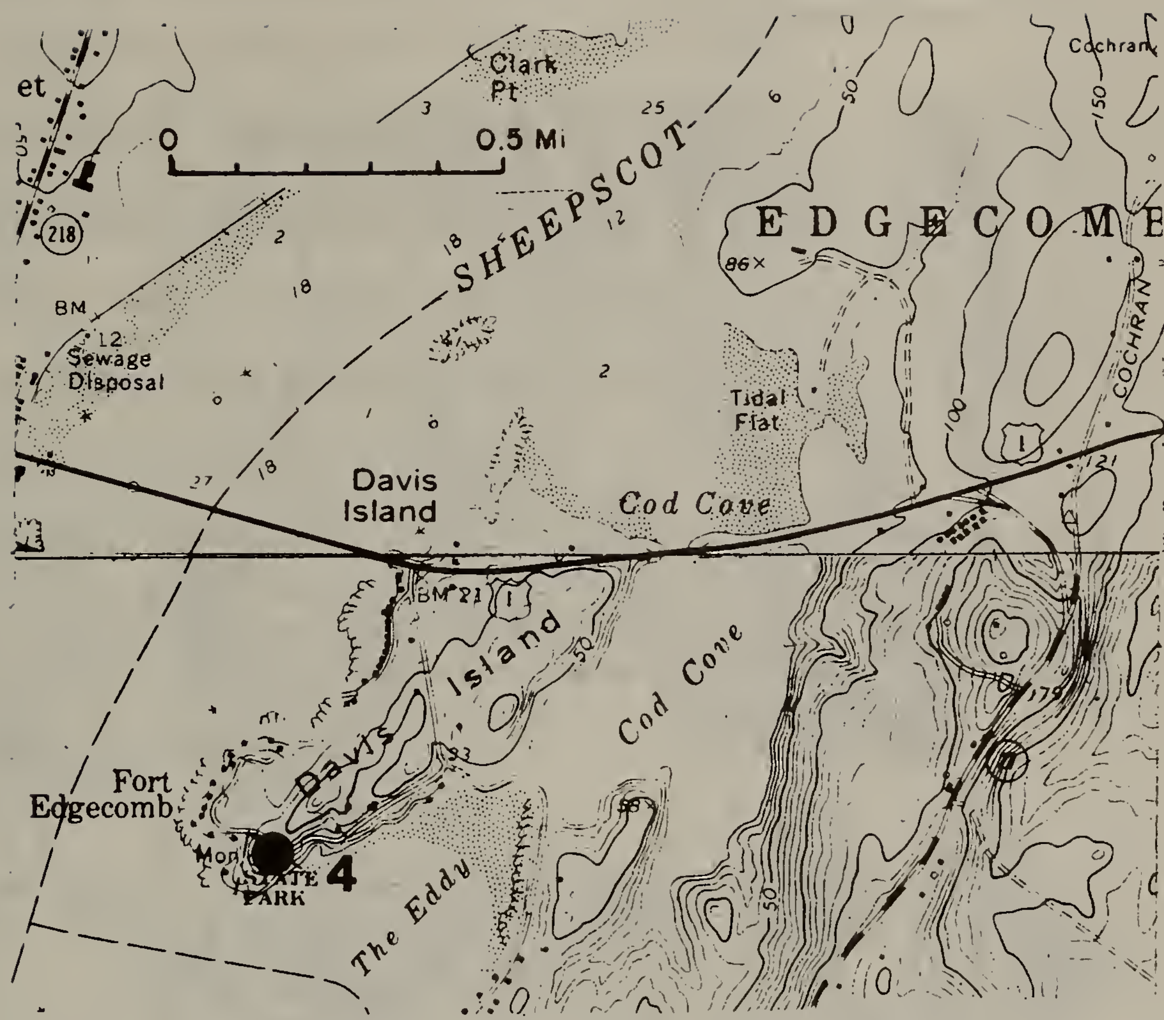


Figure 6. Location of stop 4 (Westport Island 7.5' quadrangle).

Return to cars and retrace route to Route 1

72.4 Turn left off Fort Road at yield sign.

72.8 Turn right onto Route 1.

73.7 Turn right onto Route 27. Roadcuts on Route 1 opposite this turnoff expose the contact of the Bucksport Formation with the Edgecomb Gneiss, a Devonian (?) age orthogneiss of original quartz-diorite to granodiorite composition. It is essentially sill-like and was intruded between the Cape Elizabeth and Bucksport Formations.

78.4 Boothbay-Edgecomb town line.

78.5 Park on right shoulder of Route 27 adjacent to rusty-weathering road cuts.

STOP 5 (Figure 7. Note: the locations of stops 5 through 9 are shown on Figure 8, a geo-

logical map of the central part of the Boothbay 15′ quadrangle).

Figure 7. Location of stop 5 (Westport Island and Bristol 7.5′ quadrangles).

This is an exposure of the lower member of the Cross River Formation (referred to informally by the writer in a guidebook for a Geological Society of Maine field trip in 1984 as the Pemaquid Harbor Formation, a name and useage now abandoned). The lower member here is migmatized very sulfidic mica schist, locally with abundant graphite and sillimanite. Thin interbeds of micaceous quartzite are sporadically present. This is one of the least migmatized parts of the Formation. More typically it is an agmatite with feldspar megacrysts and rafts of non-rusty quartz + plagioclase + biotite granofels and occasional calc-silicate granofels. This formation is migmatized to about the same degree that the Cape Elizabeth Formation in this area, and thus there does not seem to be a metamorphic grade difference between the Cross River Formation and the Cape Elizabeth and Bucksport Formations.

Continue on Route 27.

81.4 Boothbay Railroad Village on left.

83.4 Boothbay Center Village

83.8 Boothbay/Boothbay Harbor town line.

84.9 Turn left on Route 96 at stoplight.

86.2 Boothbay Harbor/Boothbay town line.

87.3 East Boothbay village

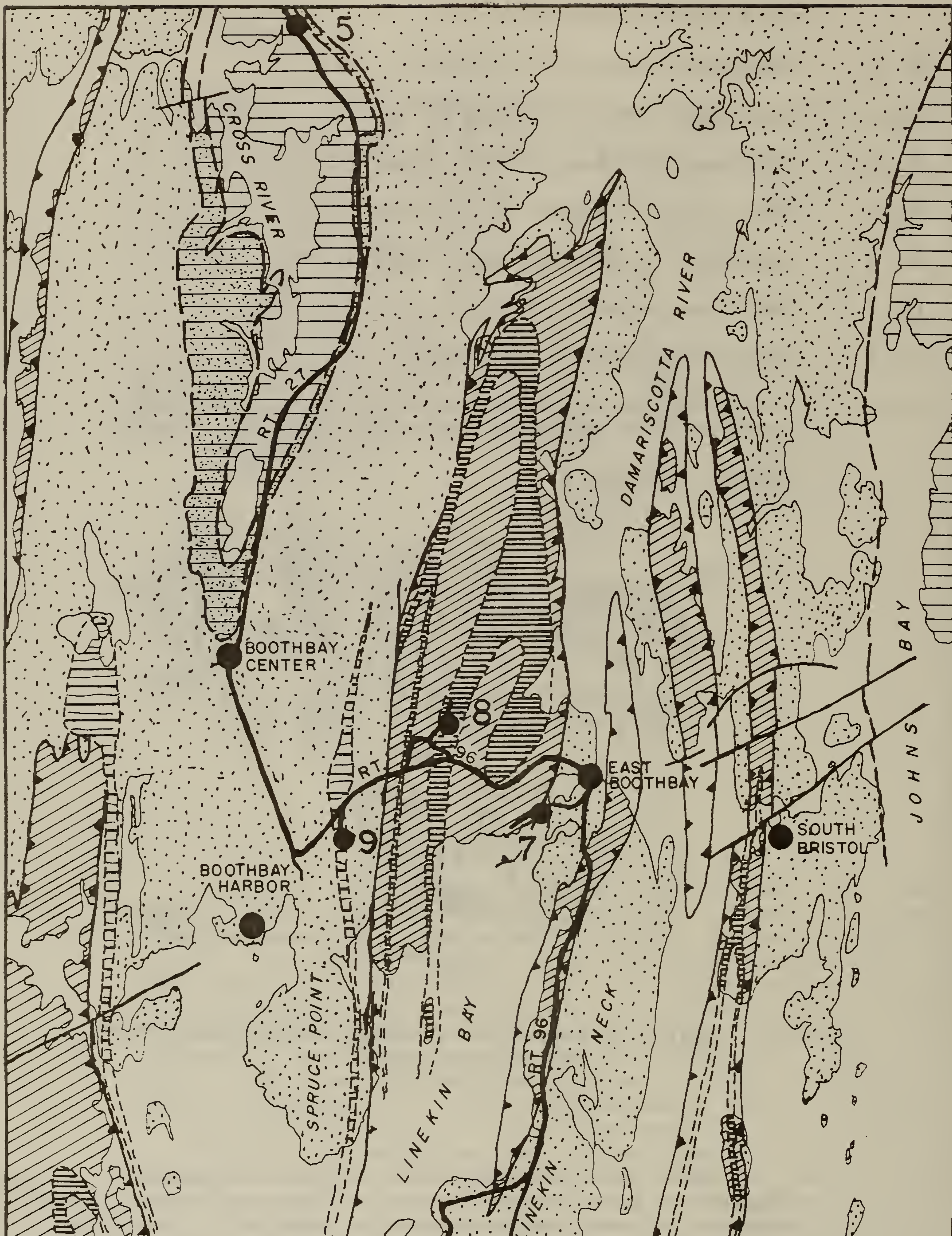
89.8 Smugglers Cove on the right.

90.9 Turn left into public parking lot. Walk down road to cove and then easterly along the shore.

STOP 6 (Figure 9)

Figure 9. Location of stop 6 (Pemaquid Point 7.5′ quadrangle).

This is probably the finest exposure of the Bucksport in the general area. It consists of brownish biotite + hornblende granofels with thin interbeds of greenish calc-silicate, and minor rusty-weathering metapelite in zones 1 to 2 meters thick. Note the abundance of asymmetric parasitic folds with frequent plunge reversals. These folds are correlated with the



5
CROSS RIVER
RT 27
BOOTHBAY CENTER
8
RT 96
EAST BOOTHBAY
7
9
BOOTHBAY HARBOR
DAMARISCOTTA RIVER
SOUTH BRISTOL
JOHNS BAY
SPRUCE POINT
LINEKIN BAY
NECK
RT 96
LINEKIN
6
0
1 mi
0
1 km

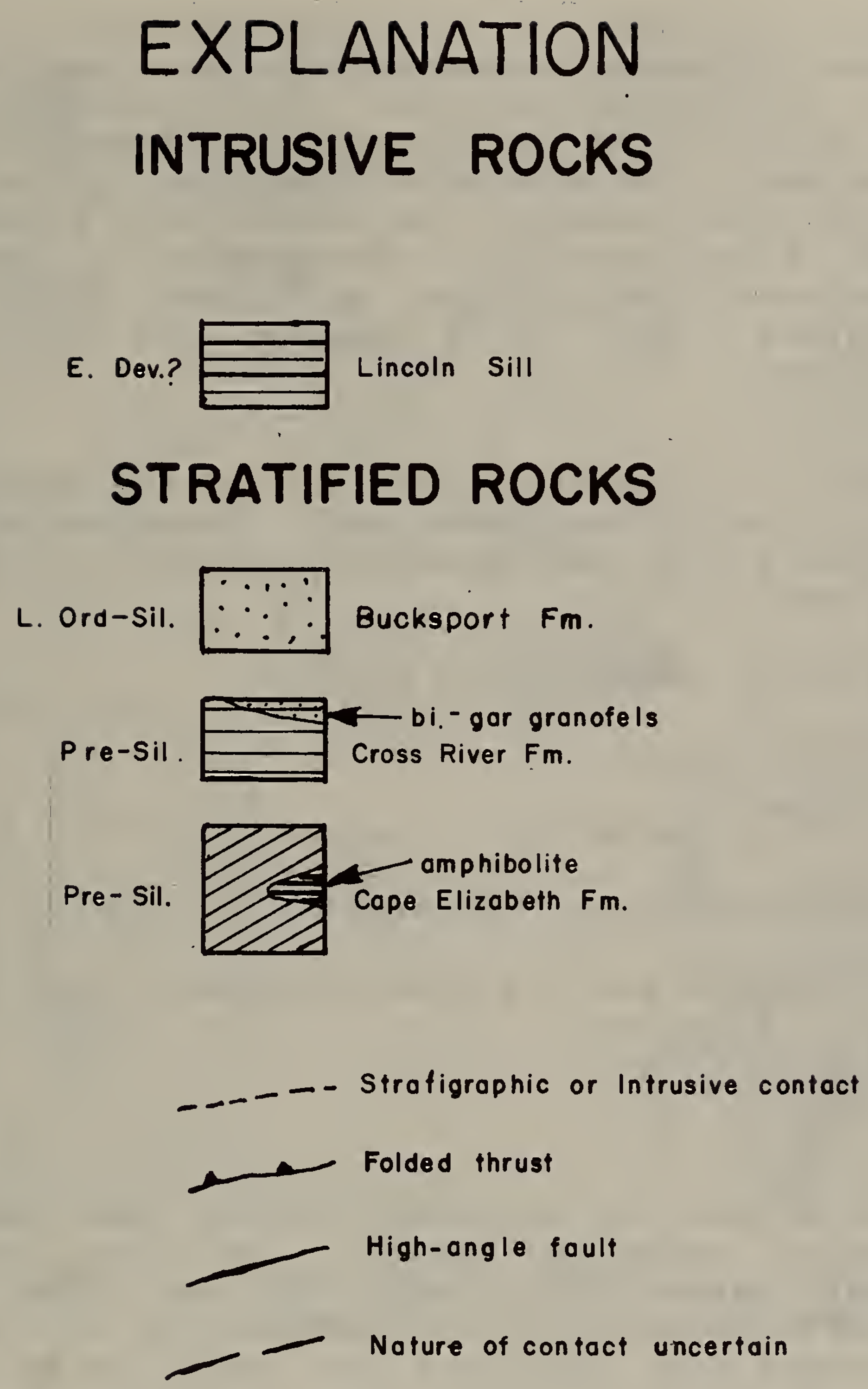


Figure 8. Geologic map of the central part of the Boothbay 15′ quadrangle. Geology by A. M. Hussey II.

F2 fold sequence of the Casco Bay Group to the west. Cleavage is weakly to moderately developed locally, but absent in most of the area. This cleavage is parallel to F2 axial planes. Although large pegmatite dikes are common here, the Bucksport Formation is not migmatized like the Cape Elizabeth and Cross River Formations are nearby. This is due to a difference in composition of the Bucksport that favors a higher partial melting temperature for it compared to the other two formations.

Return to cars and turn left out of the parking area, following the shore road. Exposures along the shore are all of the Bucksport and associated pegmatites.

91.3 Bear left at road fork.

92.3 Sharp right curve.

93.0 Turn left on Route 96.

95.7 Turn left on Murray Hill Road.

95.9 Turn left into parking area for public boat ramp.

STOP 7 (Figure 10).

Walk west (to the right facing the water) along the shore below the seawall. Up to the first dock to the west (about 500′ west of the boat ramp) the Bucksport crops out. In this portion of the Bucksport there are a few intervals of very rusty weathering sillimanite-rich schist. The contact between the Cape Elizabeth and Bucksport Formations occurs 15′ east of the dock, and here the contact appears to be a conformable one, with the Bucksport structurally over the Cape Elizabeth. Just west of the dock the Cape Elizabeth Formation includes two or three brownish-weathering skarn-like calc-silicate beds.

This contact is interpreted (but with considerable hesitation and equivocation) as a premetamorphic fault that has been folded by the F2 fold sequence on the basis of the inferred differences in age of the juxtaposed units: Precambrian for the Cape Elizabeth, and Ordovician to Devonian for the Bucksport. Although the Bucksport appears structurally over the Cape Elizabeth here, the regional interpretation is that the Cape Elizabeth structurally overlies the Bucksport. One of the questions for discussion at this locality is whether this contact, as relatively clean as it is,

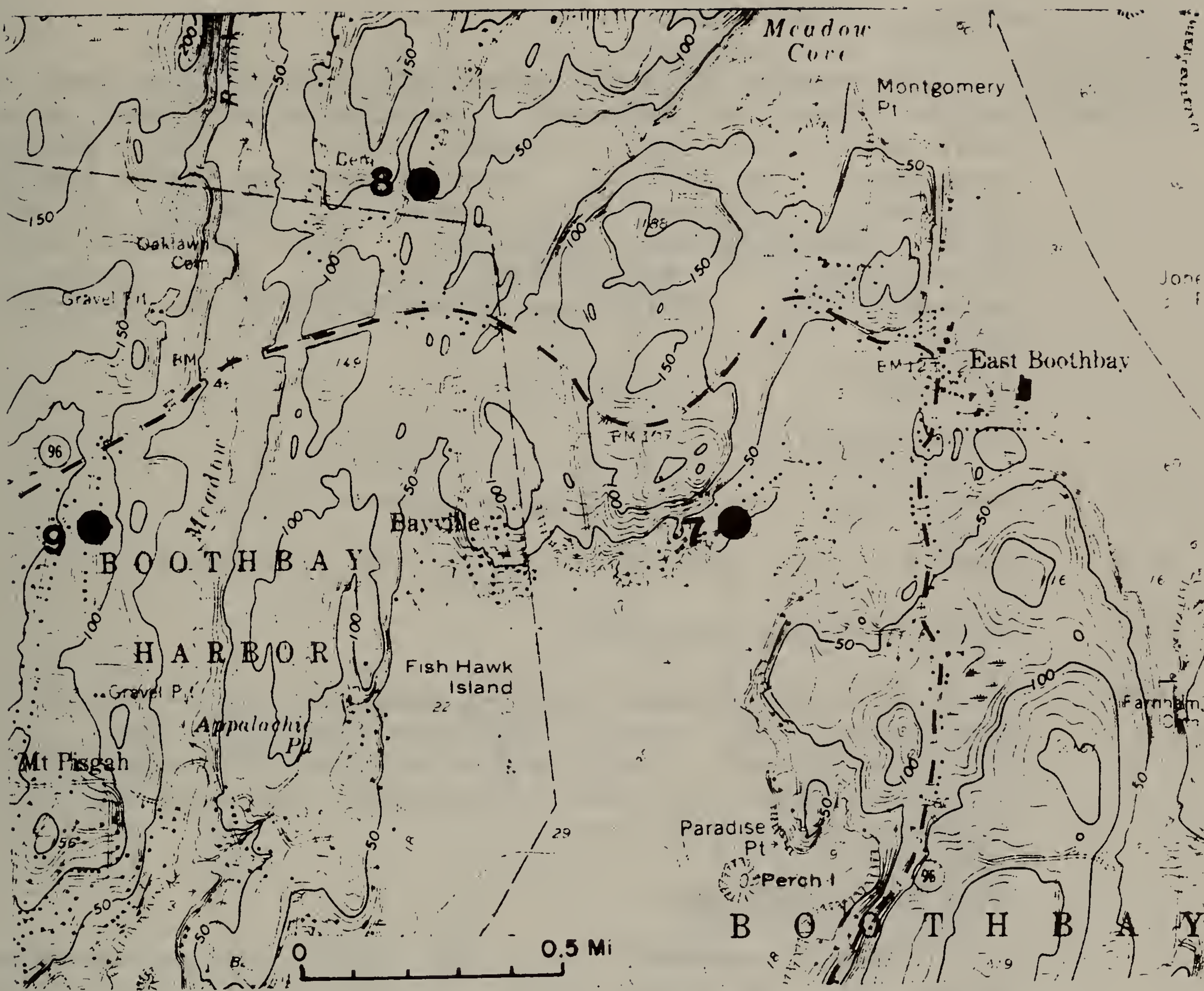


Figure 10. Location of stops 7, 8, and 9
(Pemaquid Point 7.5′ quadrangle).

could represent a mechanical surface (perhaps a wet-sediment gravity slide) despite essentially no indication of milling of the two units.

Return to cars, turn around and exit right onto Murray Hill road.

96.3 Turn left on Route 96.

97.3 Amphibolite in the nose of an F2 anticline.

97.7 Turn right onto Bradley Road.

97.9 Turn right onto Back Narrows Road.

98.1 Park on shoulder of road.

<u>STOP 8</u> (Figure 10)

Exposures of amphibolite are abundant through the woods east of the road. This is interpreted to be a unit within the Cape Elizabeth, but it may possibly represent the Spring Point Formation above the Cape Elizabeth. Although mostly massive, locally it shows thin compositional layering. This amphibolite is interpreted to have been basaltic ash prior to metamorphism. Note good examples of south-plunging F2 folds.

Turn around.

98.3 Bear right, staying on Back Narrows Road.

98.4 Road to the right. Stay on Back Narrows Road.

98.7 Turn right on Route 96.

100.1 Turn left on unnamed road at brow of hill. Park on left about 300 feet to the south, opposite the Colonial Market. Walk up the narrow paved road to the left about 300 feet to pavement outcrops on either side of the road.

<u>STOP 9</u> (Figure 10).

Glacial pavement outcrops expose the metamorphosed phase of the Lincoln Sill, a syntectonic intrusive of porphyritic mafic syenite. In the Liberty area, where it is not metamorphosed, the original igneous mineralogy consists of the orthoclase (phenocrysts and groundmass), biotite, augite, and orthopyroxene. At this stop, where the sill is thin it has been metamorphosed to a schist consisting of biotite, hornblende, and orthoclase feldspar. The orthoclase phenocrysts are aligned parallel to the foliation. The Lincoln Sill in the Boothbay Harbor area is wholly within the outcrop belt of the Bucksport Formation and has been deformed around the southern end of the Boothbay antiform. this part of the sill is discontinuous with, but lithically identical to the sill in the type area in Liberty, Maine. Acadian-age pegmatites and granite post-date the sill.

Return to the cars and turn around. Turn left on Route 96.

100.6 Turn right on Route 27, retracing route back to Route 1.

111.8 Turn right on Route 1.

112.4 Park on shoulder of Route as far off the pavement as possible. THIS IS A BUSY THOROFARE AND NOISY. THIS IS ONE OF THE AREAS WHERE SPEEDERS FLAUNT THEIR DISREGARD FOR THE POSTED 55. USE CARE IN CROSSING THE HIGHWAY!

STOP 10 (Figure 11).

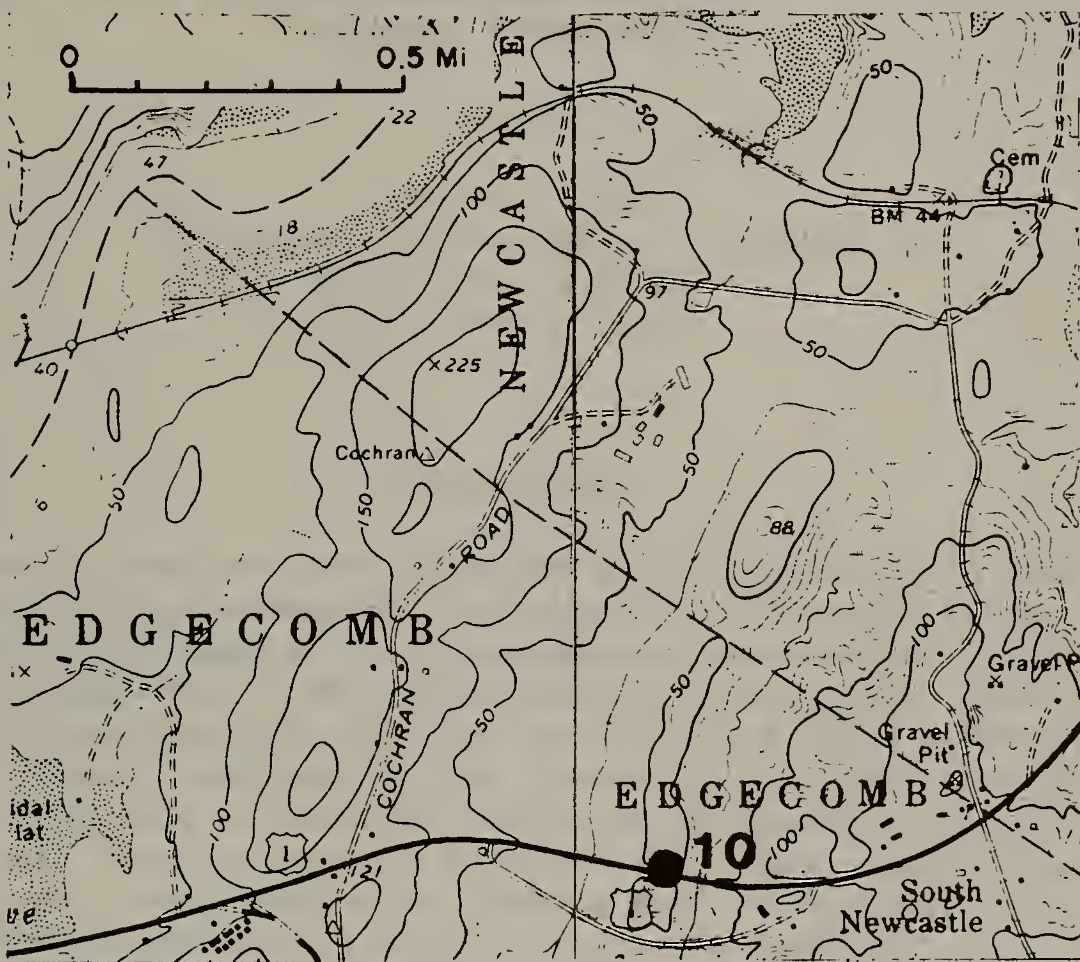


Figure 11. Location of Stop 10 (Wiscasset and Damariscotta 7.5′ quadrangles).

This is a roadcut in the Bucksport Formation, here consisting of biotite + hornblende granofels with thin interbeds of greenish calc-silicate granofels. Pegmatite and quartz pods and stringers are common. Bedding and the minor pegmatitic and quartz striners are deformed by upright F2 folds. These are cut by unfolded but slightly foliated 1 to 3 m thick pegmatite dikes. One f2 fold is transected along its axial plane by a metamorphosed but non-migmatized dark gray hornblende + biotite granofels dike. A late brittle vertical fault with prominent gouge zone (which is eroding to produce a notable vertical reentrant in the roadcut) trends N 50 E across the exposures. Slickensides at the edge of the fault suggest predominant strike slip movement. The amount of movement is indeterminant due to lack of offset marker

structures.

This is the end of the trip. Return via Route 1 to Brunswick, Route 201 to Topsham, Route 196 to Lewiston, and local streets to the Bates Campus.

TRIP B-5

MARINE GEOLOGY OF CASCO BAY AND ITS MARGIN

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INTRODUCTION

The complex bedrock skeleton of Maine's coast exercises a first order influence on regional geomorphology and permits subdivision of the coast into four morphologic compartments: the southwestern Arcuate Embayments; the southcentral Indented Shoreline; the northcentral Island-Bay Complex; and the northeastern Cliffed Shoreline (Kelley, 1986). This field trip will focus on Casco Bay, the largest embayment of the Indented Shoreline. Within this area glaciation has exercised a secondary influence on coastal geology by mantling the area with till and glaciomarine sediment which have been eroded during late Quaternary sea level changes to produce most of the intertidal and subtidal sediment of the Bay. Finally, within the past 200 years, human activity has affected the Casco Bay area by clearing forests and building river dams and coastal structures which have made more, then less sediment available to the coastal region.

The purpose of this field trip is to examine the coastal geomorphology and Quaternary stratigraphy of the margins of Casco Bay, and in conjunction with observations on the Bay's offshore geology, to consider the late Quaternary geologic history of the area, and the significance of recent human activity on its coastline.

LOCATION

Casco Bay is the first major coastal re-entrant north of Boston (Figure 1). It is 24 km by 10 km in area and bordered by Cape Elizabeth and Bailey Island on its outer southeast and northeast corners, and Portland and Freeport to the southwest and northwest, respectively.

The Bay's geometry is determined by the northeast strike of the Precambrian to lower Paleozoic rocks of the Casco Bay Group (Osberg et al., 1985). Erosion resistant ridges of these metamorphic rocks form a series of peninsulas to the northeast which become submerged ledges or islands to the

Figure 1. Location of seismic tracklines and field stops in Casco Bay.

southwest. The four prominent bedrock ridges encountered from east to west are dominated by: 1) Orrs-Bailey Island, Halfway Rock, Cod Ledges; 2) Harpswell Neck, Cliff and Jewell Islands, Green Island Ledge; 3) Mere Point, Birch and Goose Islands, Chebeague, Long, and Peaks Islands; and 4) Wolf Neck, Cousins, and Mackworth Islands.

The deepest water in the Bay lies in the Sounds between the bedrock peninsulas. Much of the inner Bay is relatively flat-bottomed between the rock islands and shoals, whereas the outer Bay displays highly irregular bathymetry (Figure 2). Deep channels extend through breaks in the bedrock ridges at several places and permit free exchange with the Gulf of Maine and a 3.75 m tidal range. Some of these channels extend toward rivers which enter the Bay, but others do not and owe their existence to pre-glacial fluvial action, as discussed below.

QUATERNARY GEOLOGY

The recent Surficial Geologic Map of Maine (Thompson and Borns, 1985) compiles all Quaternary mapping in the Casco Bay area and provides a complete bibliography on work published from the region. The record of Quaternary geology in Maine begins about 13,500 years ago with the deposition of till and stratified material in the well-described coastal moraines in Cutler and Kennebunk (Stuiver and Borns, 1975). Around Casco Bay, till deposits are mapped with an orientation which mimics the trend of bedrock (Thompson and Borns, 1985). Offshore seismic data (Figure 3) similarly show that the thickest deposits of till are generally parallel with the strike of bedrock (Figure 4).

Following retreat of the ice, glaciomarine sediment, the Presumpscot Formation (Bloom, 1960) was deposited on the isostatically depressed landscape. The time of deposition of the Presumpscot Formation has been bracketed between about 13,500 and 11,000 years before present. Some of the youngest dates from the glaciomarine sediment are from logs and seashells in the Portland area (Hyland and others, 1978). Generally speaking, the Presumpscot Formation is thickest in bedrock valleys and thin to absent on hill tops. This may be a result of the original deposition of the glaciomarine sediment or of reworking of the material by marine processes during emergence. Indeed, the tree fragments described by Hyland and others (1978) may have been emplaced in a slump from the Western Promenade of Portland into the ancestral Fore River valley during emergence.

The main result of the thick accumulation of glaciomarine sediment in bedrock valleys is the derangement of Maine's stream drainage. As recorded in bridge borings (Figure 5), the Fore and Cousins Rivers, though small in discharge, possess some of the greatest bedrock valleys in Casco Bay today, but are choked with sediments. The Presumpscot and Royal Rivers have relatively large discharge today, but pass over water falls before entering the ocean and are not in pre-glacial valleys. It is likely that the ancestral Kennebec/Androscoggin River even entered Casco Bay through Maquoit or Middle Bay, but is so deranged by glacial sediment today that not even a small stream marks the trace of the old valley. Kelley and others (1986) have explained the muddy nature of Casco Bay, as contrasted with the sandy environment of

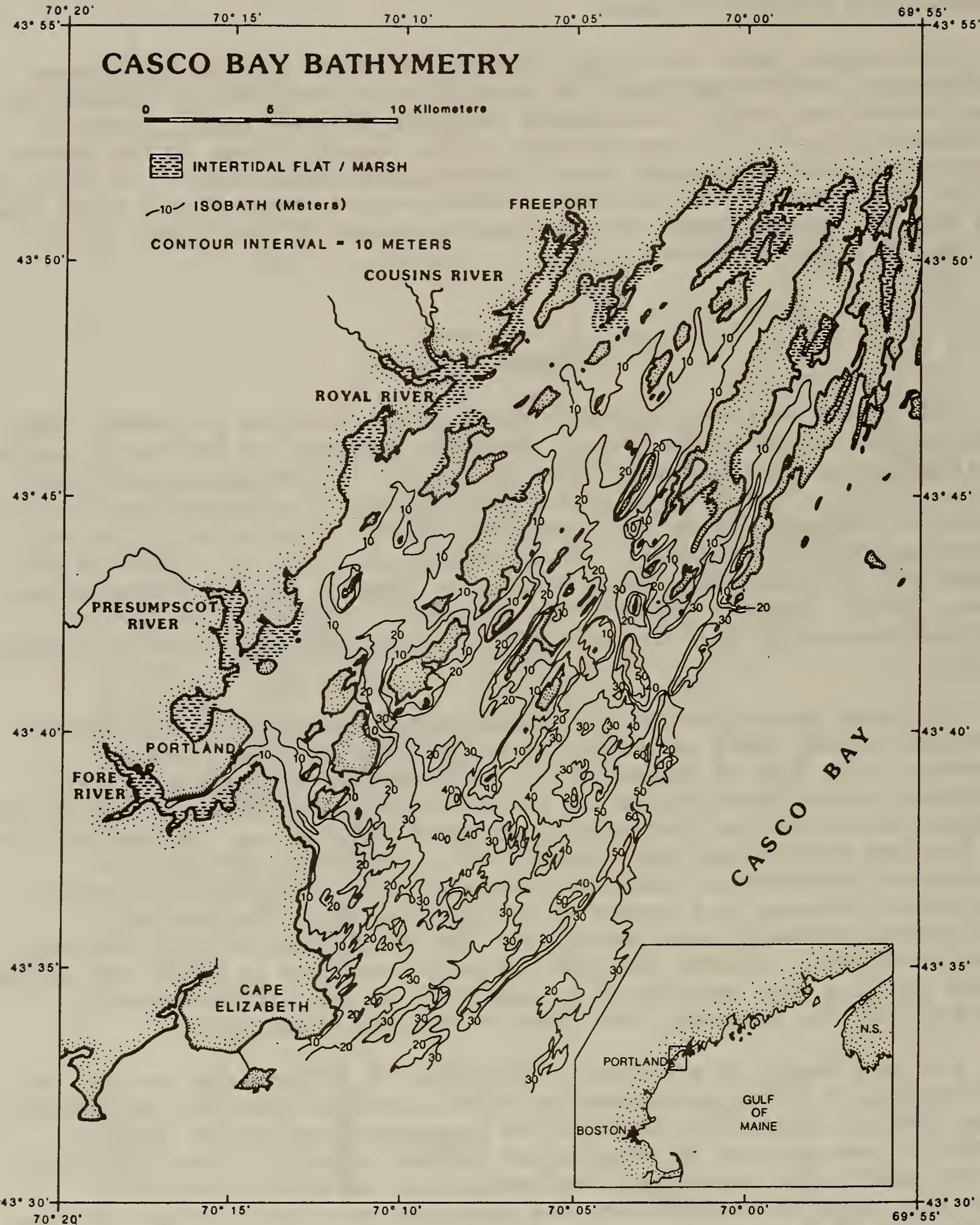

Figure 2. Bathymetry of Casco Bay.

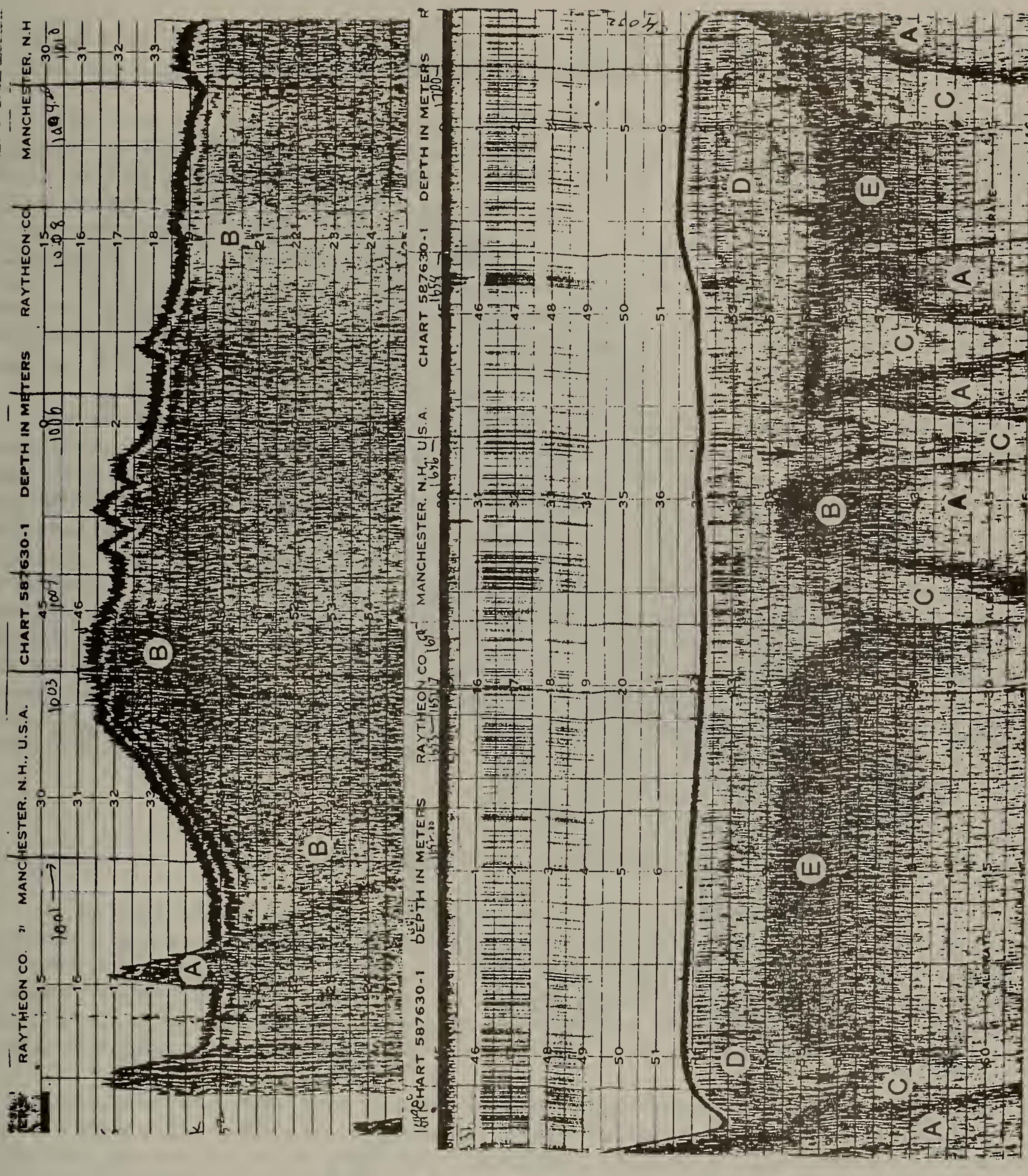

Figure 3. Original seismic records from Lines 15 (above) and 36 (below). Locations of the lines are shown in Figure 1. The interpretation of the seismic units has been described by Kelley and others (1986) and Belknap and others (1986) and is as follows: A) bedrock; B) till; C) glaciomarine sediment (Presumpscot Formation); D) Holocene mud; E) natural gas. Vertical scale is 1 m per horizontal lines; horizontal scale is about 600 m. Arrows point to inferred unconformity at surface of glaciomarine sediment.

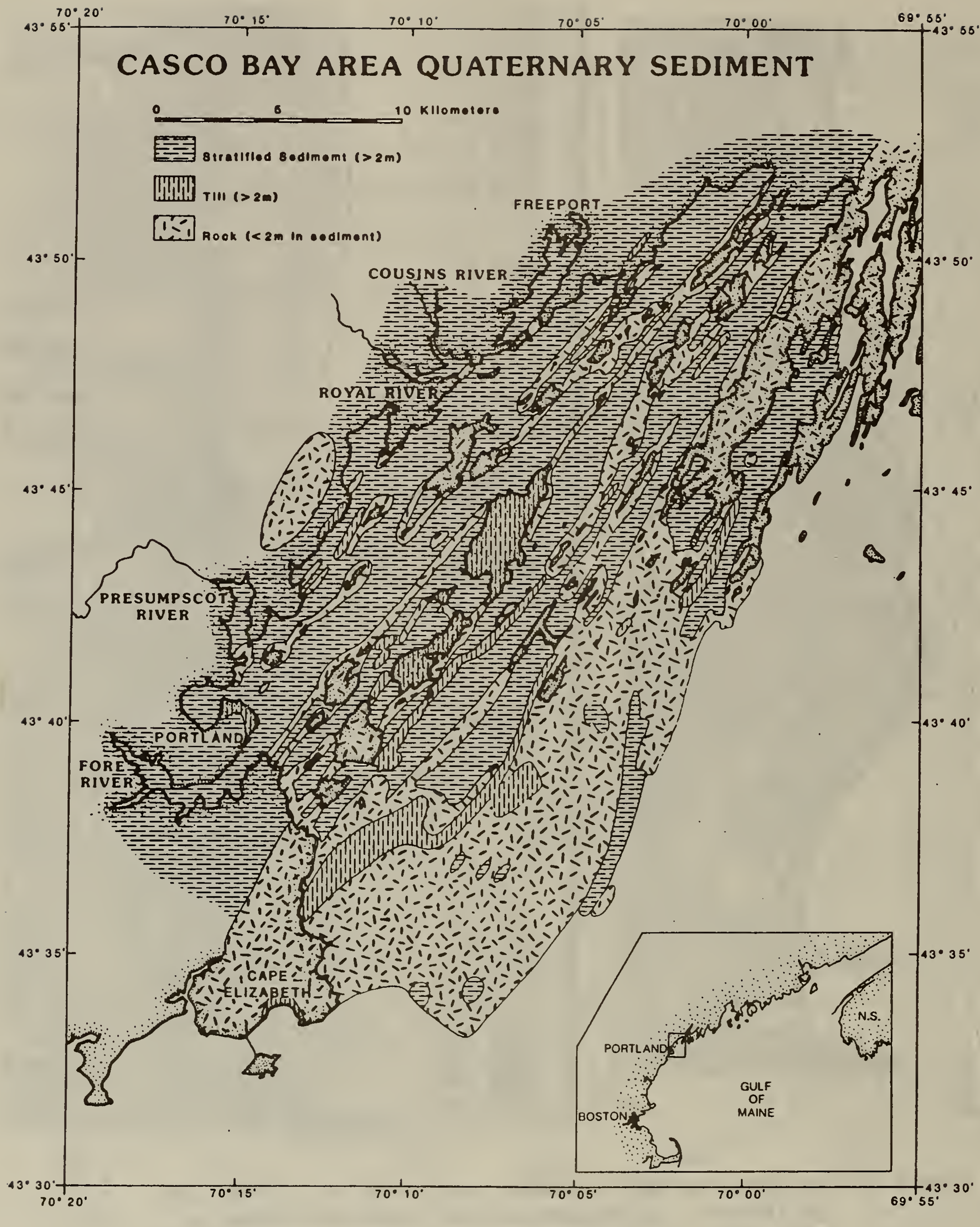

Figure 4. Surficial maps of Casco Bay based on seismic profiles.

Saco Bay, as a result of the deranged, low gradient streams entering Casco Bay versus the high gradient Saco River to the south.

The time of the sea level lowstand is not known with certainty, but is assumed to have occurred 9,000 years ago. There are better constraints on the depth of the lowstand, and several reports suggest that it was around 65 m below present sea level (Kelley and others, 1986; Belknap and others, 1986). While there is no such direct evidence of the lowstand for Casco Bay, features which appear to be of fluvial origin appear near that depth at Halfway Rock in Casco Bay (Figure 6, CB-6). The rivers which formed during the lowstand of sea level were immature streams cutting down onto a former seafloor. Since this was a muddy substrate, many of the streams were gullies at their head. Some of these gullies will be visited in the Bunganuc Bluff stop.

As sea level rose across outer Casco Bay area, it eroded sediment from most bedrock pedestals and left a coarse sediment pond in adjacent topographic lows (Figure 6). Thus much of the outer Bay is exposed rock or coarse sediment (till or lay deposits). Mud exists mostly in well protected paleovalleys (Figures 6, 4). As the gullies formed during the lowstand were drowned, they became productive estuaries partly filled with organic matter. After burial with modern mud, this organic matter has evolved into natural gas deposits, which trace the evolution of valleys from the lowstand to the inner Bay (Figure 7). The inner Bay has a more complete stratigraphic column than the outer Bay because it has more recently come under the influence of marine processes and is sheltered from the ocean by numerous islands (Figure 7). Nevertheless, even in the inner Bay, the smooth seafloor is being eroded in some places by tidal current scour.

HUMAN ACTIVITIES ON THE COAST

A general model for the evolution of Maine's embayments has been presented by Kelley (1986). Casco Bay fits this model (Figure 8) well with an outer wave eroded area of bare rock and coarse beaches, central region of eroding bluffs and mud flats, and an inner area of salt marshes and aggrading mud flats. The form of this model moves landward as sea level rises, leaving a coarse rocky inner shelf across most of the Gulf of Maine.

Human activity has profoundly affected this coast, and interacted with its evolution for the past 200 years. Early colonists cut numerous trees along the shoreline which subsequently eroded because of the loss of vegetative cover. As a result, harbors like Mast Landing and Wharton Point became useless for commerce and are today salt marshes. Most of the Holocene sediment in Maquoit Bay is probably younger than 200 years although as yet no dates are in hand (Figure 9). Following deforestation, vegetation began to stabilize slopes again and dams were built on rivers trapping sediment. Thus, coastal areas which had been blanketed with an influx of sediment were slowly cut off from new sand and mud. As a result, most of the marshes which fringe many of the bluffs in the inner Bay owe their origin <u>not</u> to muddy rivers, but to episodic slumping (Figure 10).

The most important human influence on Casco Bay today is in the form of residential house construction. Ignorant of the long term evolution of Casco

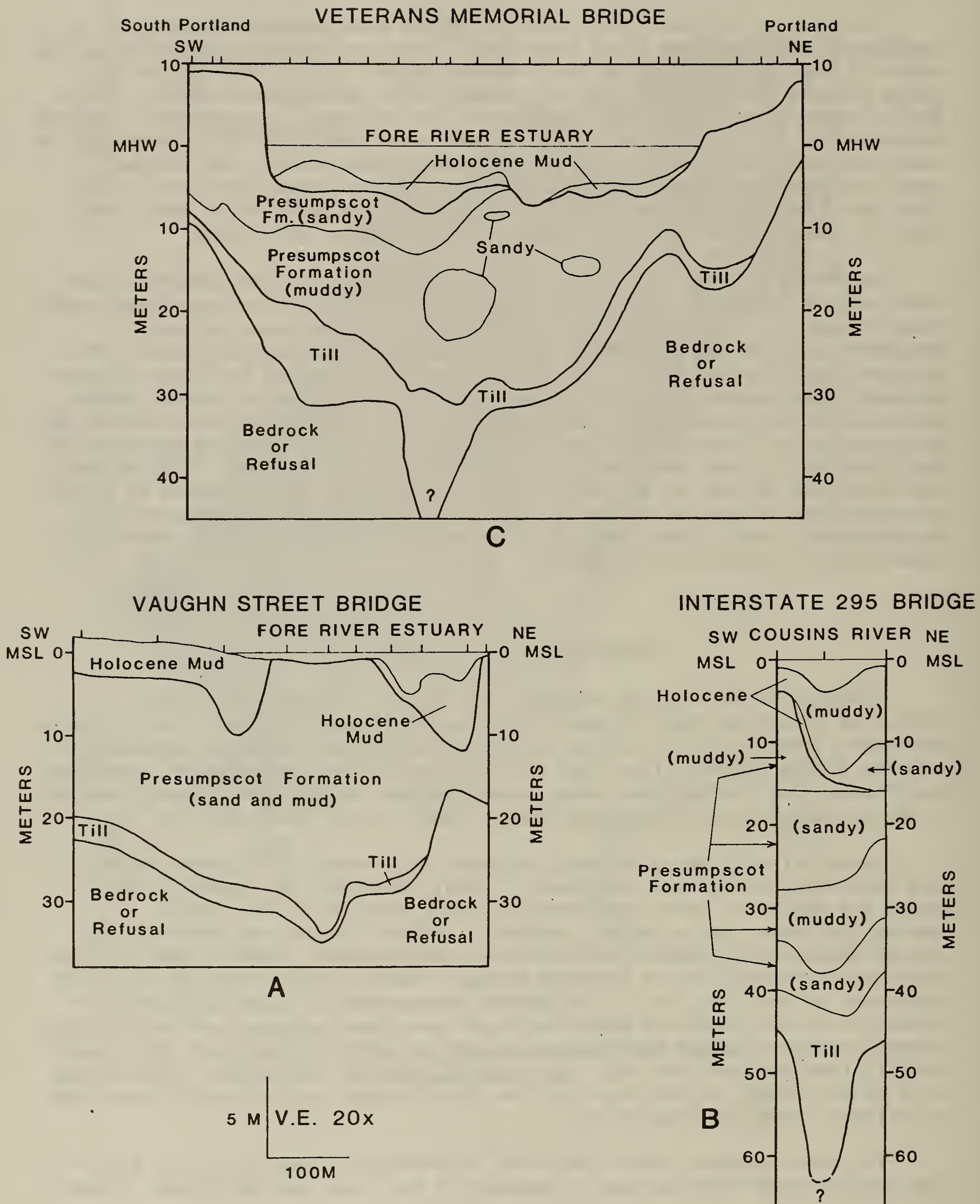

Figure 5. Logs of valley fill in Casco Bay's major river valleys.

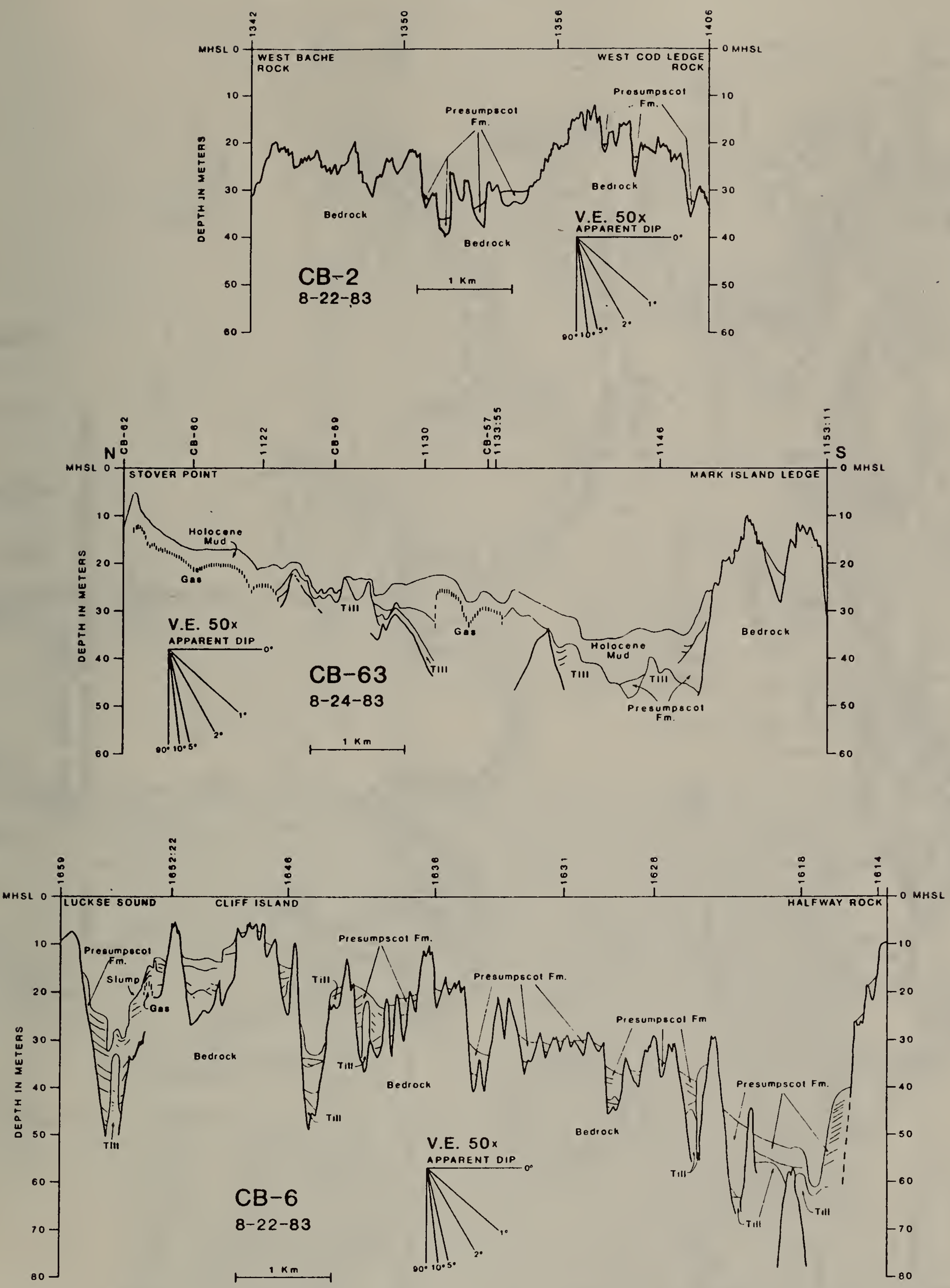

Figure 6. Interpreted seismic profiles from outer Casco Bay.

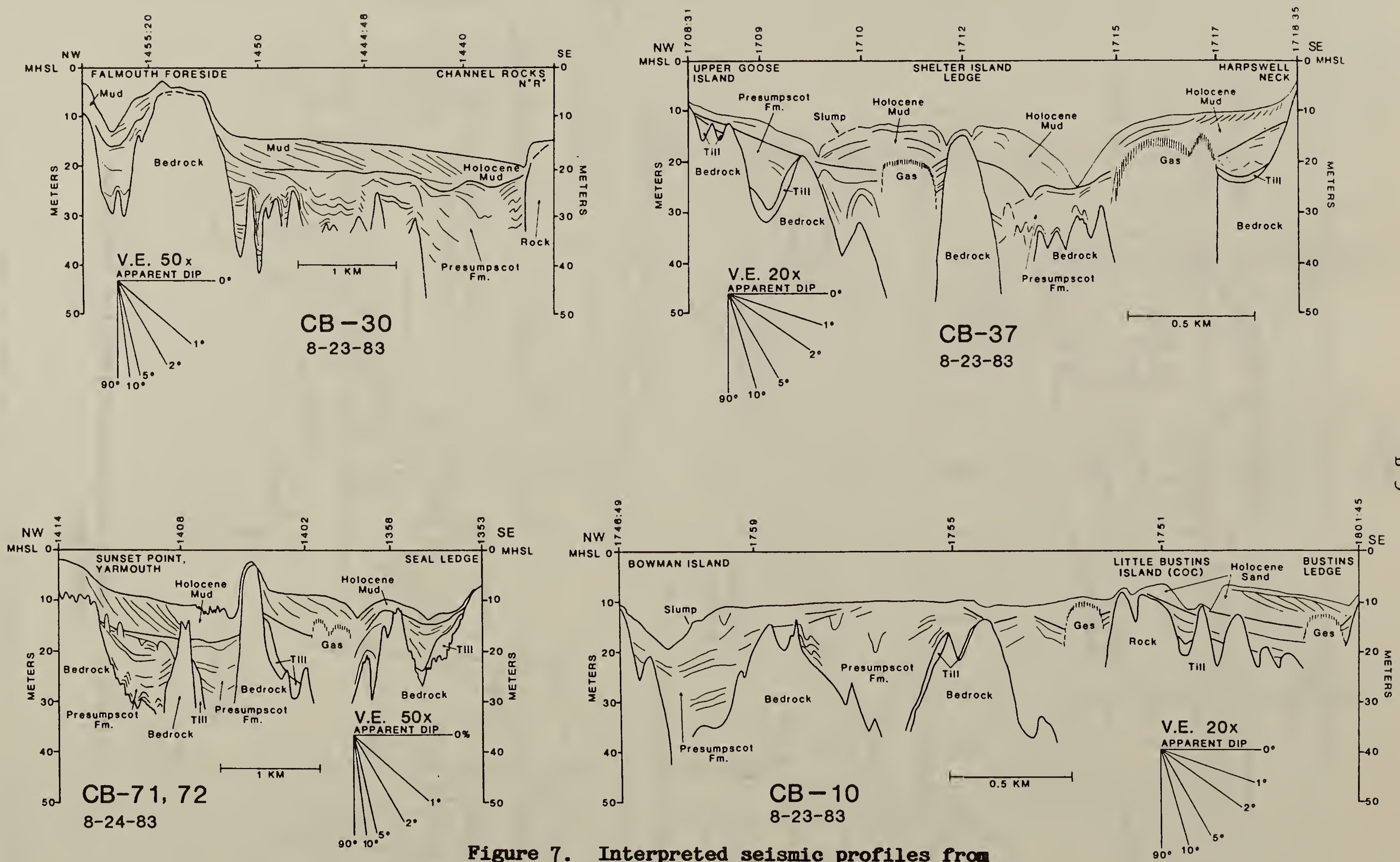

Figure 7. Interpreted seismic profiles from inner Casco Bay.

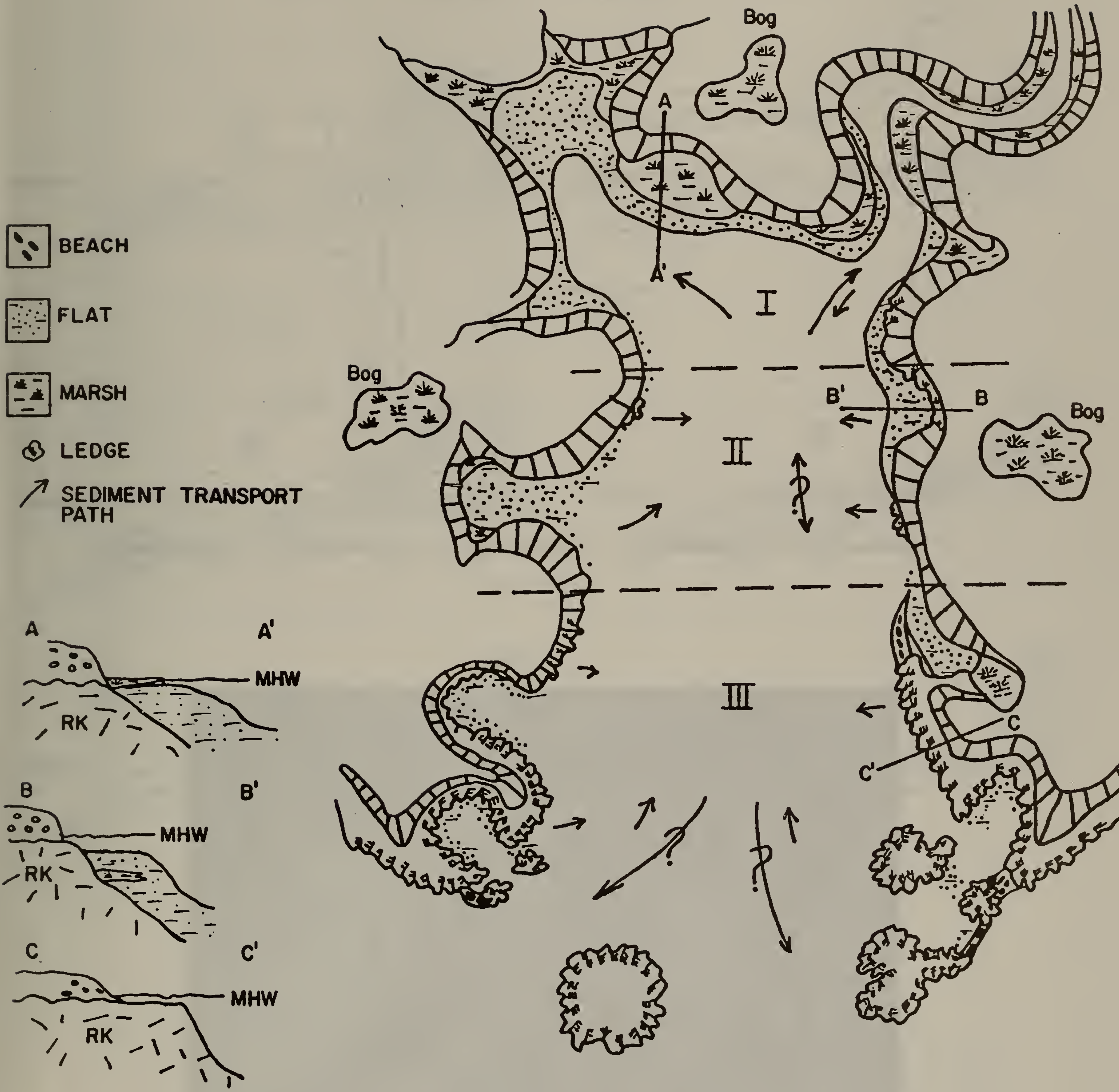

Figure 8. A general model for the evolution of an ideal Maine embayment.

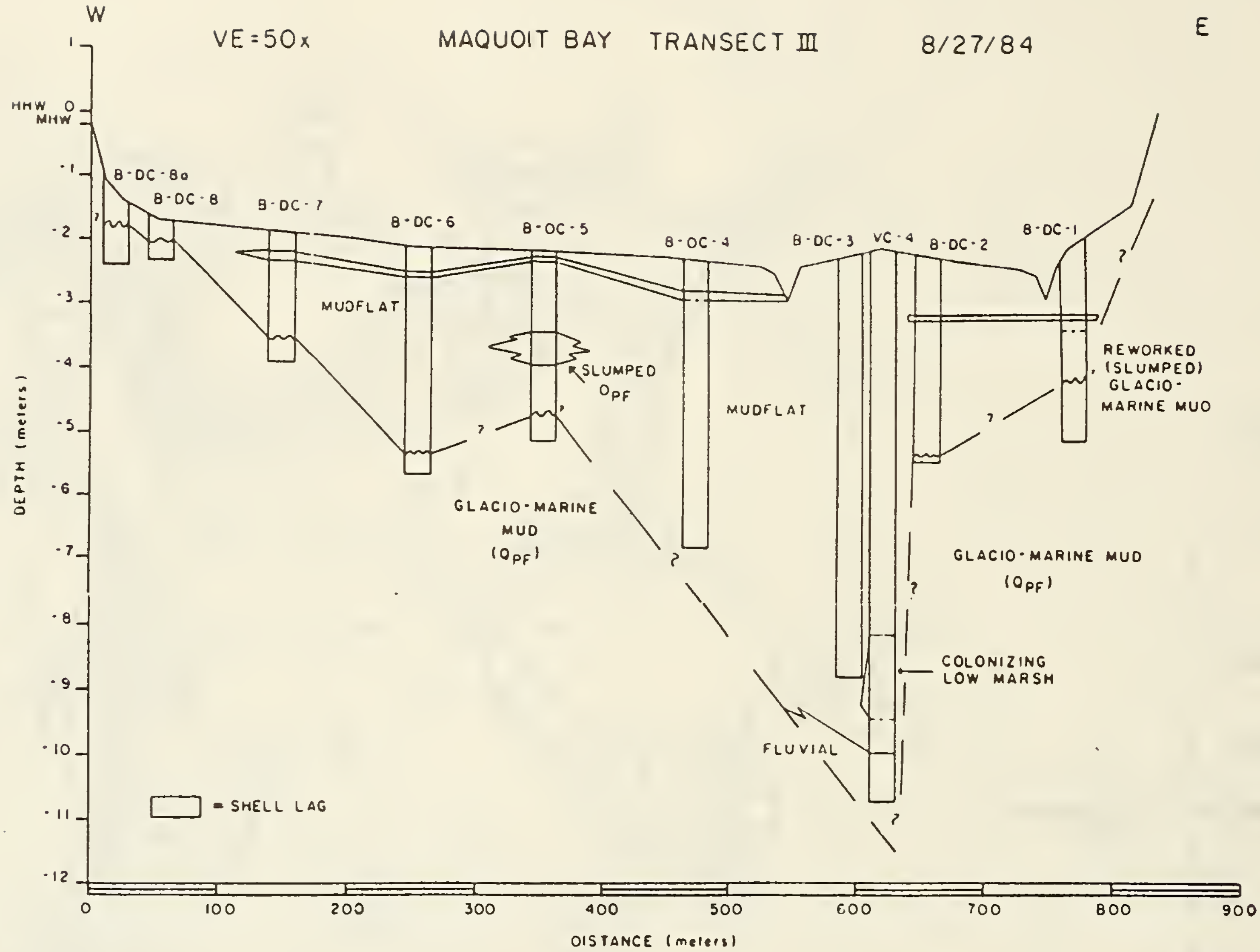

Figure 9. Cross section of Maquoit Bay (upper). Photo showing gullies and eroding Bunganuc Bluff (lower).

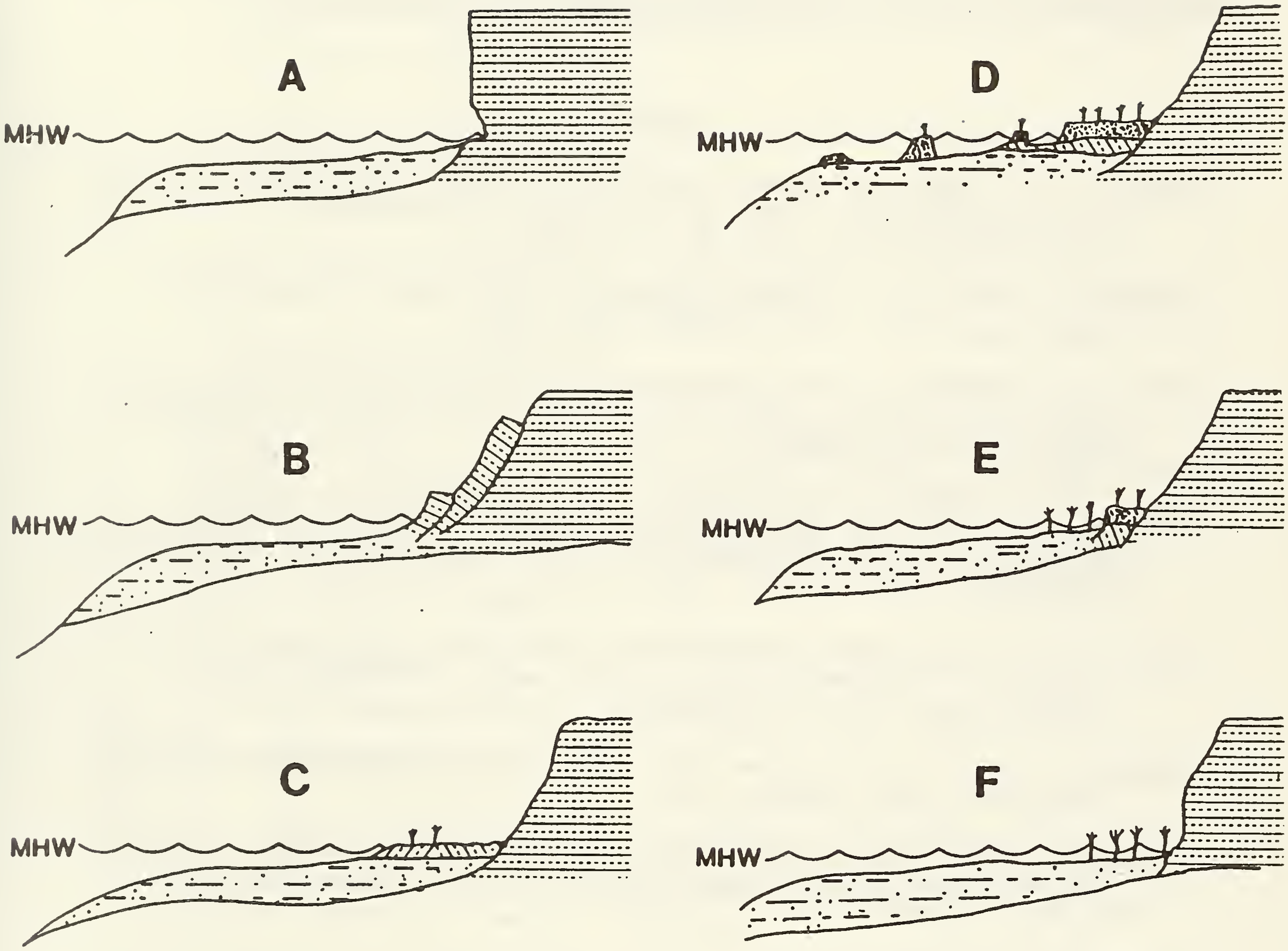

Figure 10. A model for the formation of Maine's fringing marshes by slumping bluffs.

Bay, homeowners are increasingly building houses at the edge of bluffs of the Presumpscot Formation. When they recognize the dynamic nature of the bluffs (Figure 10) they frequently attempt to "stabilize" them with bulkheads. In the short term this robs the intertidal mud flats of much-needed mud, in the long term it is unlikely to halt the inexorable rise in the ocean.

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ITINERARY

Assembly point is at Two Lights State Park in Cape Elizabeth, Maine. This is most easily reached by taking Route 77 south from Portland, and then taking Two Lights Road.

Mileage

0.0 STOP 1: At Two Lights State Park--we will briefly examine the high energy shoreline here. Unconsolidated sediment is present only as gravel beach deposits between resistant ridges of bedrock. Because bedrock strikes to the northeast, Casco Bay's beaches almost exclusively face the northeast or southeast.

Return to cars and drive north on Two Lights Road.

0.5 Note the flat sandy soils of this area. Much of this sand was probably left as regressive shoreline deposits.

1.0 Turn right onto Route 77 north.

1.6 As we drive perpendicular to the bedrock trend the road crosses numerous ridges and valleys.

2.6 Turn right onto Shore Road.

3.9 Note at this rare ocean view in Cape Elizabeth, that gravel beaches dominate the outer Bay's intertidal areas.

5.0 Turn right into Fort Williams.

5.5 STOP 2: Park at Portland Head parking lot. In this high energy coastal setting in the outer Bay, only coarse grained beaches protected by bedrock ridges can endure in intertidal environments. Thus, structures built high on rock ledges are the only coastal structures which can survive. As the ocean drowns this area the beach deposits will become analogous to the seaward dipping ponds of sediment on seismic line CB-6 (Figure 6).

Return to cars.

6.0 Turn right onto Shore Road (Cottage Road).

7.9 Turn right onto Broadway.

8.5 Note views of Portland which are favored for condominium construction now. The question facing the public is whether the loss of intertidal access and environments is worth the tax base provided by the high density housing proposed for this area.

9.0 Turn right onto Pickett Street.

9.2 Cross Fort Street onto Fort Preble.

9.3 STOP 3. Park at Old Settlers Cemetary. Willard Beach is the largest beach in Casco Bay. Like the others we have seen it faces the northeast, but is protected by nearby islands.

Recreational growth in this area recently prompted the City of South Portland to request the Army Corps of Engineers nourish this beach. The Army agreed to increase the width of the beach, which is presently eroding more than one foot per year, but at a cost of $800,000 plus $100,000 per year thereafter. The City would be required to pay 50% of this. The erosion problem at Willard is partly a result of small seawalls built by homeowners. The cost of the project would be picked up by the Army if the same homeowners were not so close to the project site. Thus human action exacerbates the beach erosion, and the

presence of beach houses precludes correction of the problem, because the City cannot afford the project.

Return to cars.

9.5 Cross Fort Street and drive down Pickett Street.

9.6 Turn left on Broadway.

10.3 Turn right onto Cottage Drive which becomes Main Street.

11.2 Cross the Million Dollar Bridge. Figure 5 shows borings from Fore River bridges. This was once a major river valley entering Casco Bay. Bear right after crossing bridge onto Route 1A.

12.0 Bear right onto Fore Street. This newly renovated area is the Old Port section of Portland.

12.8 Cross India Street and proceed up the Eastern Promanade. This low area was once Clay Cove, the original settlement site of Portland in the 17th century.

13.3 STOP 4: Fort Allen Park. This site, which is a hill of glacial till, provides a fine overlook to discuss the seismic stratigraphy of the inner Bay (Figure 7). The impact of human filling is evident along the harbor which was once a mud flat. Condominiums now compete with fishermen for space along the waterfront.

Return to cars and turn right onto Eastern Promanade.

14.3 STOP 5: Back Cove. This is a brief stop to examine the stratigraphy of Back Cove as revealed by borings for the new bridges. It seems likely that the Presumpscot River once passed through Back Cove toward the Fore River and carved the wave cut cliffs and terraces of the area when it did. Human filling is the major process here today, and the Cove is less than half its original size. The Cove is clean now, however, and a wildlife sanctuary. The City used to dump raw sewage into the cove.

Continue down hill.

14.5 Turn right onto Washington Avenue. Bear right onto I-95 at end of bridge.

15.3 Turn right onto Route 1, Falmouth.

16.0 Cross Presumpscot River mouth. Mackworth Island is visible offshore.

17.1 Turn left onto Old Route 1.

17.2 STOP 6: Turn left onto Gilsland Farm and park in Maine Audubon lot. We will walk out the path toward the river. At the top of the flat hill it is clear we are standing on the flat, former seafloor. To either side marshes fill in former gullies which were once the site of

numerous slumps. We will examine the eroding bluff facing the river. Note how ineffective the dead vegetation is in preventing erosion...in fact it leads to more erosion by slowing down new plant growth. The bluff here is composed of sand and mud layers typical of the regressive phase of the phase of the Presumpscot Formation. Note the eroding marshes. Even in the muddiest river in Maine, sea level is rising faster than mud can build new marshes.

Return to cars and drive to Route 1.

17.5 Turn left onto Route 1.

19.2 Turn left onto Route I-95 access.

19.4 Turn right onto I-95.

24.9 Cross Royal River. This is another deranged stream with a waterfall on the left side of the road. Attempts to maintain a harbor where the muddy freshwater flocculates upon entering the sea are futile, but costly.

26.6 Cross the Cousins River. Note the esker at the left. This is an old river valley (Figure 5) choked with glacial sediment.

29.3 Turn right into Freeport.

30.5 Turn right onto Bow Street

31.5 Mast Landing. Note this was once the site where tall masts were exported to Europe. It is now filled with marsh. Note the numerous Y turns where long masts were moved by oxen to the sea. The logs precluded right angle turns.

32.3 Note gully on right. This was cut into the flat former seafloor during a lower sea level.

32.9 Bear left <u>away from</u> Wolf Neck.

34.0 Turn left onto Flying Point Road.

36.6 Bunganuc Landing. Another mast landing area in an old gully.

36.7 Turn right onto private dirt road. This turn is just <u>before</u> the left turn onto Casco Road.

37.4 <u>STOP 7</u>: Bunganuc Bluff at end of dirt road. Travel down clammers path along gully wall. Note the size of the gully. There are many in the area. They lead to channels out in Maquoit Bay (Figure 9) which in turn lead to gas filled valleys in inner Casco Bay (Figure 7). The bluffs here are among the highest built by the Presumpscot Formation. This is the ice-proximal, transgressive phase of the glaciomarine sediment and numerous load structures may be seen.

Evident here also is the problem of new house construction on bluffs. If marshes are to continue to benefit our clam flats (Figure 10), these bluffs must erode.

Return to cars.

38.0 Turn right onto Flying Point Road. Bear right onto Bunganuc Road immediately.

39.8 Wharton Point. Another mast landing site that is filled in. Turn left up sandy hill. This is sandy, regressive sediment which blocked the old Kennebec River from entering Casco Bay.

41.6 Turn right at stop sign onto Mere Point Road.

42.4 Bear left onto Middle Bay Road.

43.5 Turn right onto Harpswell Neck Road.

44.9 Note that at this low point salt marsh is visible at both sides of the road. Soon Harpswell will become an island.

52.9 Turn left onto Stover Cove Road. Continue on gravel road toward water (bear right, then left at Harpswell Sound). Drive onto sandy beach.

54.0 STOP 8: Stover Cove Spit. This is one of the only spits of any size in Casco Bay. It owes its origin to erosion of an enormous moraine which is mostly submerged beneath Harpswell Sound (Figures 4, 6). This spit is protected by several laws, and will never be developed.

TRIP B-6

CONTAMINANT HYDROGEOLOGY OF SOLVENTS, GASOLINE AND SALT

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INTRODUCTION

In the last several years, there has been an enormous boost to the study of hydrogeology by the unfortunate need to investigate and clean up chemical spills of one sort or another. On this trip we will be seeing (as far as it is possible to see groundwater) the nature of three very different kinds of spills, and several new and old tools that can be used to further spill investigations. We will also discuss what can be done for unfortunate owners of wells in the paths of plumes, and what lies ahead in terms of prevention and cleanup.

HYDROCARBONS AS GROUNDWATER CONTAMINANTS

Hydrocarbons have been with us for decades. The chlorinated hydrocarbons are commonly known as **solvents** because of their property of dissolving oily materials (for which water's nickname as the universal solvent is not apt). Chlorinated hydrocarbons are the quintessential degreasers, plasticisers, and paint strippers. No doubt they have been improperly disposed of since they were first manufactured, but it was only in the '70s that leaks were discovered to be causing groundwater contamination.

Nowadays, that contamination is known to be nationwide and alarmingly ubiquitous: it may have been for years, but only since about 1980 have chemical analytical techniques been able to detect hydrocarbons down to the parts per billion range.

Chlorinated hydrocarbons are manufactured by substituting a chlorine atom for a hydrogen, somewhere in the chain or ring. This may be done at one location per molecule, as in (mono)chlorobenzene, or at several, as in trichloroethylene. The result is a compound which has a greater specific gravity than its non-chlorinated cousin.

Properties of common hydrocarbons, both chlorinated and not are given in the following table:

Table 1: Some Interesting Properties of Hydrocarbons

Hydrocarbon	solubility in water mg/l (=ppm) @ 20°C	specific gravity	recommended maximum contaminant level (ppb)	odor recog. threshold (mg/cu m)
chlorobenzene	500	1.11	60	1
1,1,1-trichloroethane	4400	1.35	200	400
trichloroethylene	1100	1.46	0	110
tetrachloroethylene	140	1.63	-	50
pentachlorophenol	14	1.98	220	?
2,3,7,8-TCDD (dioxin)	0.00002	-	-	-
benzene	1780	.88	0	0.5
toluene	515	.87	2000	1
xylene	175	.86-.88	440	<1

You will notice that hydrocarbons are far from being insoluble. Some are soluble in water in the parts per thousand range, though considered as a group, their solubilities vary over several orders of magnitude. Because of the extreme insolubility of some (especially dioxin) we can be thankful that they are unlikely to be groundwater contaminants (though they can and do adsorb to soil and sediment particles).

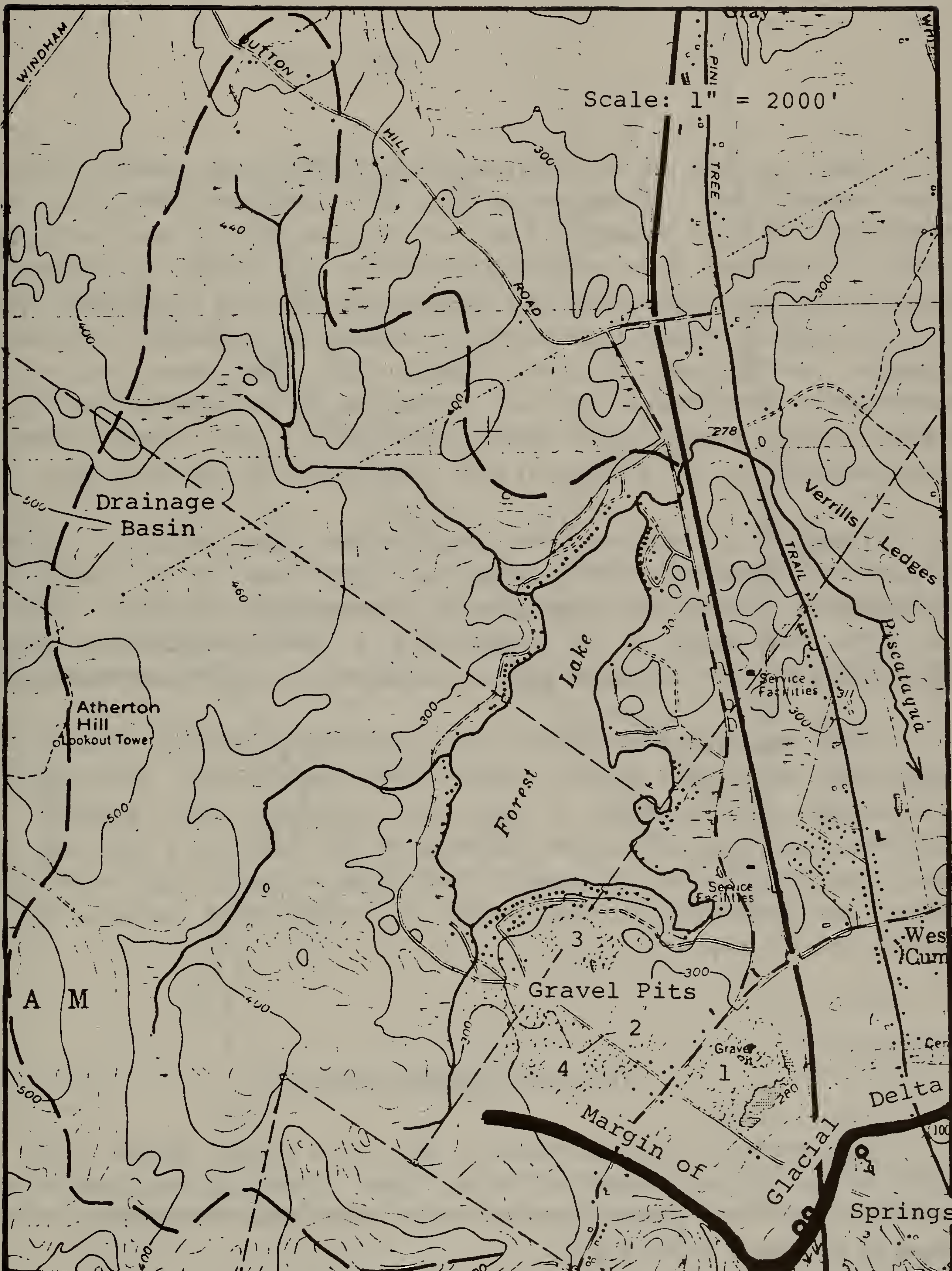


FIGURE 1. West Cumberland site with Forest Lake and its drainage basin dammed by glacial delta sands and gravels. (USGS 7½ minute Quadrangle: Cumberland Center).

Also note that all the chlorinated hydrocarbons are heavier than water, some markedly so. Because of this, and combined with their relative insolubility, they are sinkers: they tend to sink through and settle on the bottom of aquifers, from which position they are difficult or impossible to recover, and from which they can slowly leach into the groundwater passing by. The non-chlorinated hydrocarbons, including all the legion components of gasoline, are floaters in their product form. But when any of these hydrocarbons becomes a dissolved component of groundwater (and there is always some aliquot that does dissolve), it will move along with groundwater in the same direction and more or less at the same rate.

Toxicity of hydrocarbons also varies considerably. Maximum contaminant levels in drinking water have not been set for nearly all the hydrocarbons, though the table gives a representative sampling. Notice that the EPA has seen fit to recommend a zero contaminant level for trichloroethylene and benzene due to suspected or known carcinogenicity.

Odor recognition threshold is an interesting variable. It is of course subjectively dependant on the victim's nasal sensibilities. But note that it is hundreds of times easier to smell one hydrocarbon than another. Some spill sites have been discovered because of the distinctively odd odor of one minor component. The converse is that there may be plenty more cases of solvent contamination out there which lie undiscovered because the water doesn't smell funny.

THE WEST CUMBERLAND SITE

This area of West Cumberland lies on a classic glacial delta, which dams up the southern outlet of Forrest Lake. Overlying the granite bedrock are thick sand and gravel deposits which have been extensively excavated down to the water table, Fig 1.

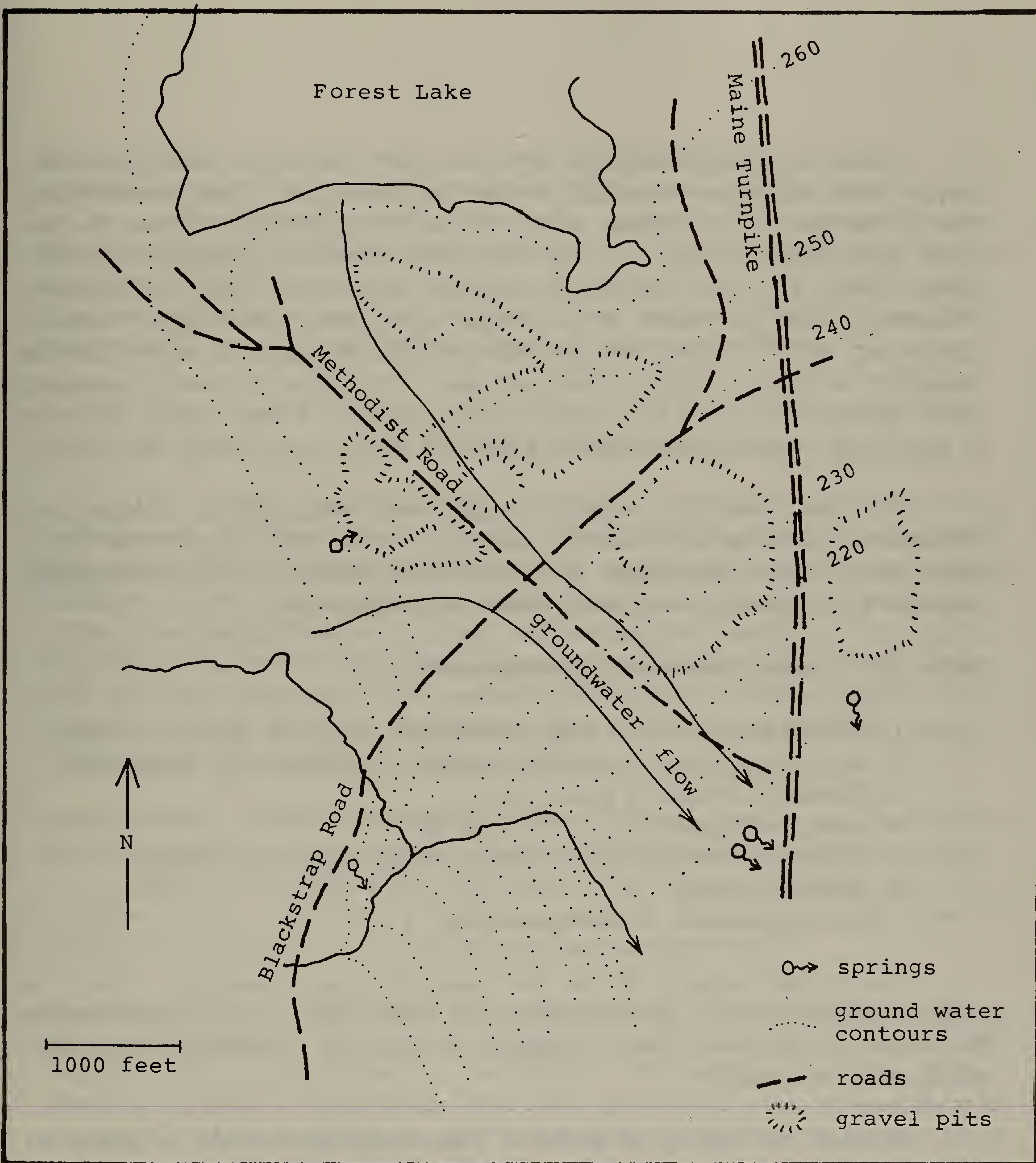


FIGURE 2. West Cumberland glacial delta with superimposed ground water contours. Flow is south-east, from the lake through the sand and gravel to the springs.

There is a surface water outlet at the north end of the lake, but some, maybe most lake water flows out through the delta dam. Early winter water levels observed in the bottom of the gravel pits provided the data for the water table map of Fig. 2. Note that water flows southwest through the glacial delta, and exits at several springs at its toe, near the turnpike. Calculating from the slope of the water table, and assuming a hydraulic conductivity of 10^{-2}cm/sec, the seepage velocity through the aquifer can be calculated at 10s to 100s of feet per day. This rate of flow is obviously much faster than that in the Sebago granite bedrock below, where fracturing is slight and characteristic bedrock yields are only a few gallons per minute.

Into this beautiful hydraulic system was introduced a contaminant, tetrachoroethylene (also known as perchlor). It is not obvious how it got there, but consider the threats to ground water listed in the following table. Perhaps it's surprising there haven't been more problems.

Table 2: Local Threats to Groundwater

1. Rinsing road tar from road construction trucks by use of solvents.
2. Auto salvage yard operations (gasoline, crankcase oil, degreasers).
3. Midnight dumping in gravel pits.
4. Leaks from gasoline or fuel oil tanks.
5. Disposal of household chemicals through septic drainfields.
6. Salting of roads
7. Dust suppression on Methodist Road

Two further aspects of this site make the case interesting. First, the problem was discovered as the result of a family feud, not primarily through the smell of perchlor, which happens to have a moderately high odor recognition threshold.

Second, the spilling of perchlor, presumably somewhere in gravel pit #4 may have taken place many years ago. It could have sunk through the sand and gravel to the top of the bedrock surface, where it continues to leach slowly into the bedrock aquifer giving the same levels of contamination in

downgradient household wells for the last three years at least. It is virtually impossible to locate the remnant pool of solvent, let alone clean it up, so it may continue to contaminate that aquifer for many years to come. While the solven t pool was sinking through the sand and gravel, it was no doubt contaminating the upper aquifer too, though only for a short period. That contaminated water has long since been flushed through: it may only have taken a year at the calculated rate of ground water flow.

THE PROBLEM WITH LUST

LUST, for the benefit of the uninitiated, is the acronym for Leaking Underground Storage Tanks, currently the sexiest topic in contaminant hydrology. Most underground storage tanks contain petroleum products, and because of the vast numbers of tanks (tens of thousands in Maine, and millions nationally) there is no quick and inexpensive solution to the problem of LUSTs. The ultimate solution is decades away and depends heavily on the level of public awareness of the problem and what can be done about it. The tools at hand to deal with the leaking tank issue are varied and complex. Some of the more significant ones include:

Identification of the location of tanks and assessment of the relative risk they pose to existing water supplies or known ground water resources. It is important to assess the risk posed by a given facility, so as to prioritize action for existing facilities, and to determine what level of precaution to take for a new or replacement storage facility.

Re-assessment of the need for underground storage facilities on a site-specific basis. Many tanks exist as a "convenience" to the owner and may not justify the risk posed by the facility. For example, the Maine DoT is removing hundred of tanks which have been determined to be non-essential to operations. Many homeowners with buried backyard heating oil storage could just as easily store their fuel in the basement.

Implementation of state-of-the-art technology for new facilities. Corrosion has been a prime cause of storage facility failures in the past. Fiberglass and corrosion-protected steel tanks can effectively deal with corrosion. Double wall tanks and dual containment storage systems can prevent future ground water contamintaion by detecting problems before they affect the environment. Training and certification of the people who must install this new technology is also important. Old skills and practices must be refurbished so that the new technology is properly installed and performs according to plan.

Formulation of a plan for existing tanks. It is neither economically nor practicalbly feasible to replace all existing storage systems overnight. While assorted early leak detection tools are available, including inventory of tank contents, ground water monitoring wells, precision tank testing, and assorted electronic monitoring devices, no method is perfect, and every method only detects a leak after it has occurred. In many cases, especially in Maine's bedrock aquifers, even a very small leak can cause very significant problems. One strategy might be to replace tanks before they leak, but try to convince a tank owner that a storage facility must be replaced even though it may not be leaking — yet! To get an idea of the range of possible options on this one issue, take a look at an EPA worksheet, reprinted as Fig. 3.

SALT AS A GROUND WATER CONTAMINANT

Salt is very soluble. Salt water is also heavy, so it sinks through the aquifer: therefore it is more likely to contaminate drilled wells than dug wells. Also it is not very toxic except for sensitive folks (who perhaps should be drinking distilled water anyway). These things make salt a very different contaminant from hydrocarbons.

In the sixties, Maine relied on the spreading of pure salt for winter highway maintenance. This salt was stored under cover, so the storage wasn't a threat to groundwater, though the spreading was. In 1968, salt use on roads reached a peak of 100,000 tons per winter: contaminated roadside

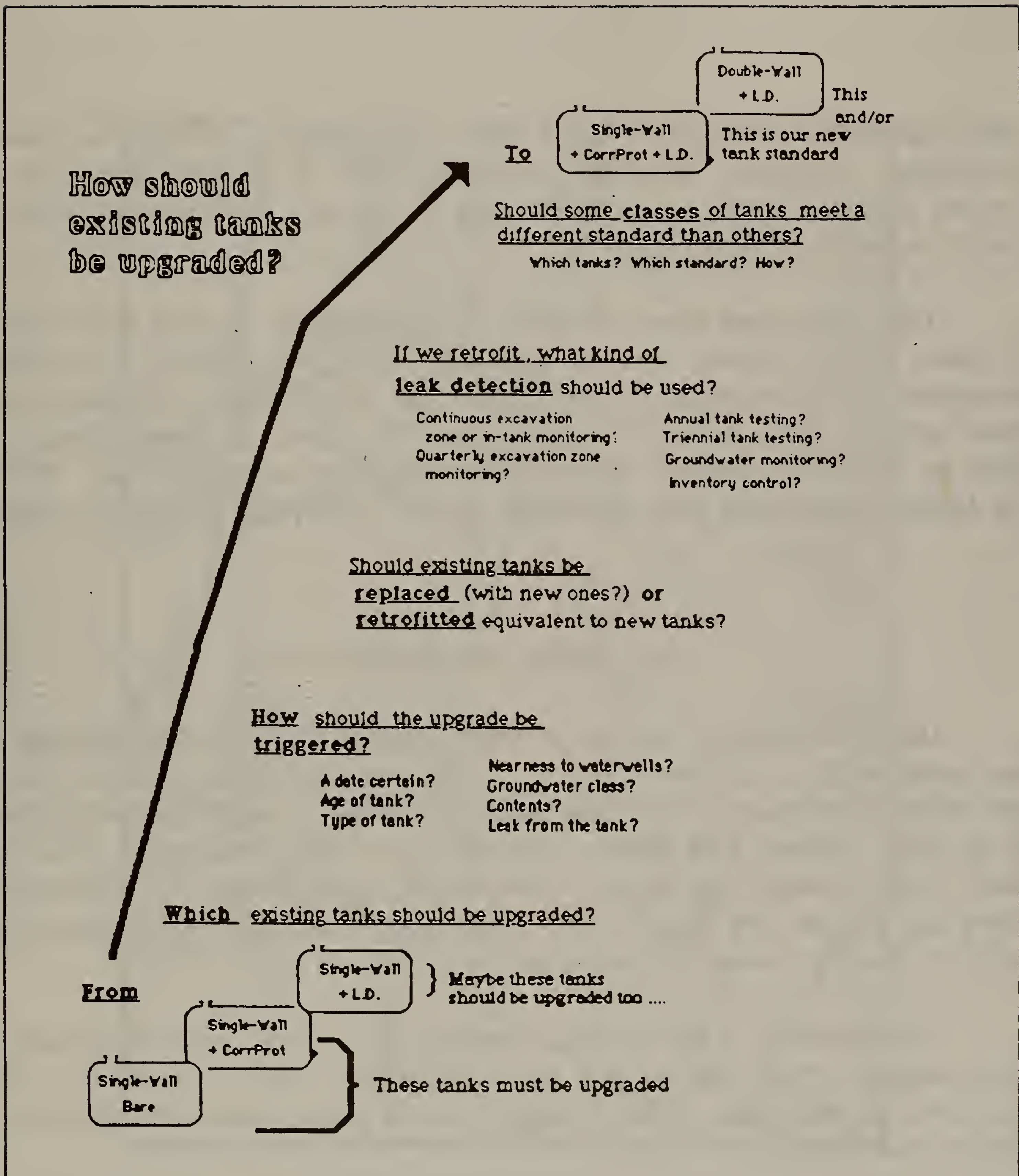


Figure 3. EPA worksheet dated 6/86, discussing what needs to be done about existing underground petroleum storage tanks.

wells reached a peak too. Since then, the use of pure salt has been nearly eliminated. Instead, sand/salt mixtures (10/1 is typical) are now used to provide traction. Now only 50,000 tons of salt are required per winter. But... how to store the huge piles?

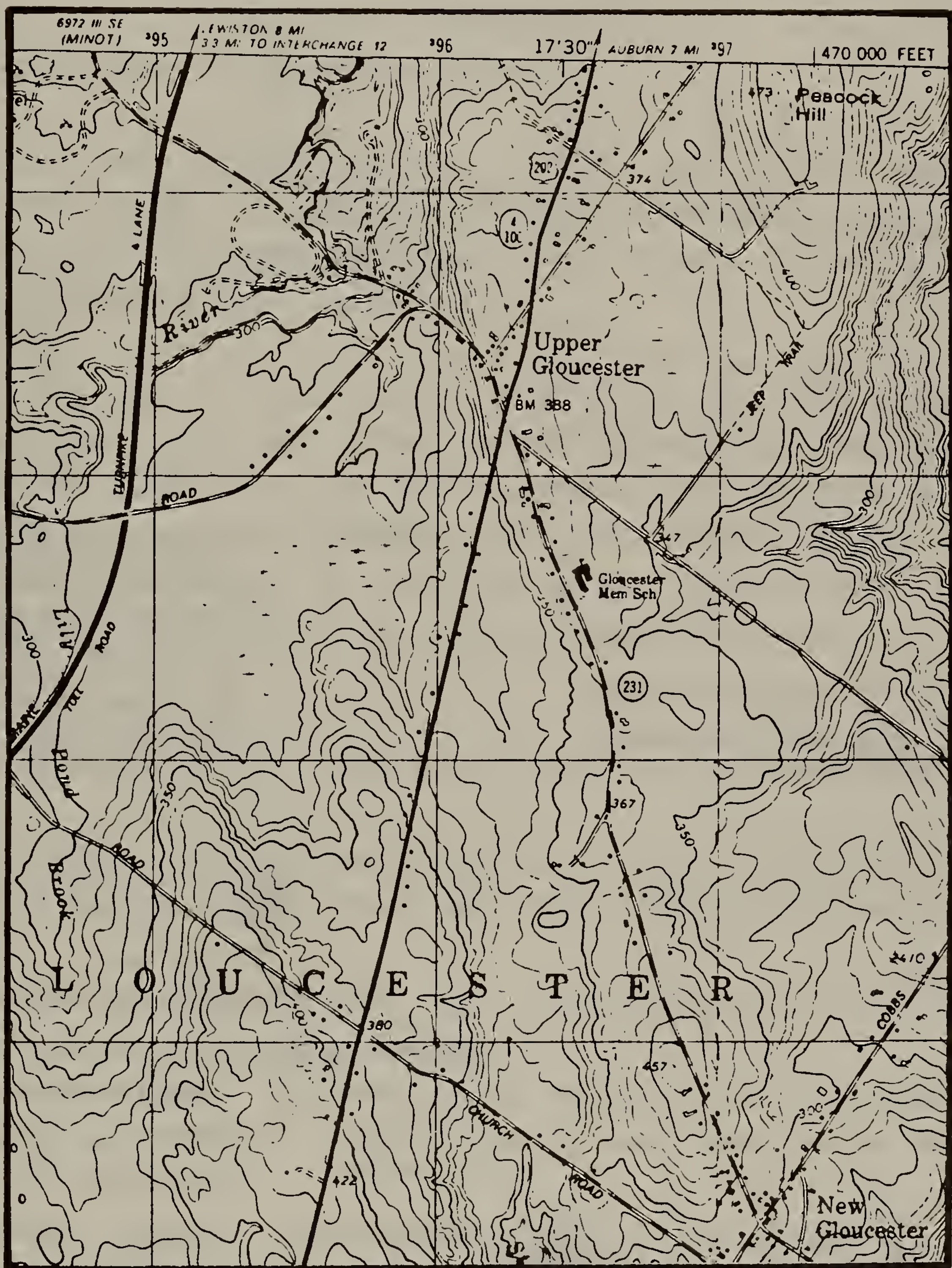
The piles have been left open to the weather, so that rain water is free to leach the salt down into the ground. Thus the problem has shifted from spreading the contamination all across the countryside to concentrating it in small areas. The solution is to cover the piles, or alternatively to move them to where the ground water is discharging to a major river. This is what the Maine Legislature has mandated for all 750± piles across the State.

THE UPPER GLOUCESTER SITE

Upper Gloucester lies on a thick basal till sheet, thick enough to show the morphology of drumlins, Fig. 4. We will be walking around the crest of one drumlin, beneath which the bedrock lies at a depth of 60-100 feet. The till is very uniform and dense. So dense in fact, that split spoon samples taken from below the water table during the drilling of monitoring wells came up dry for the most part: It is likely that such groundwater as does exist in the till moves in fractures.

The bedrock is the Sebago granite again. Not much is known about it here because there are so few local outcrops. But cores show it to be well fractured at the top: drilled wells in this area yield variously up to ten gallons per minute.

The water table in Upper Gloucester is up close to the ground surface, a fact which has allowed the development of dug wells throughout the village. But because Upper Gloucester is located on a hill top, the hydraulic gradient is predominantly downwards. This is shown schematically in Fig. 5. We will observe an astonishing 8-9 ft head difference over a 50 ft vertical spacing of monitoring well piezometers. This of course is only possible



Scale: 1" = 1300'

FIGURE 4. Topography of Upper Gloucester. The east side of the map including Route 231 is all underlain by thick basal till. Peacock Hill, Upper Gloucester ridge and the 457 ft. hill in New Gloucester are all interpreted as drumlin landforms.

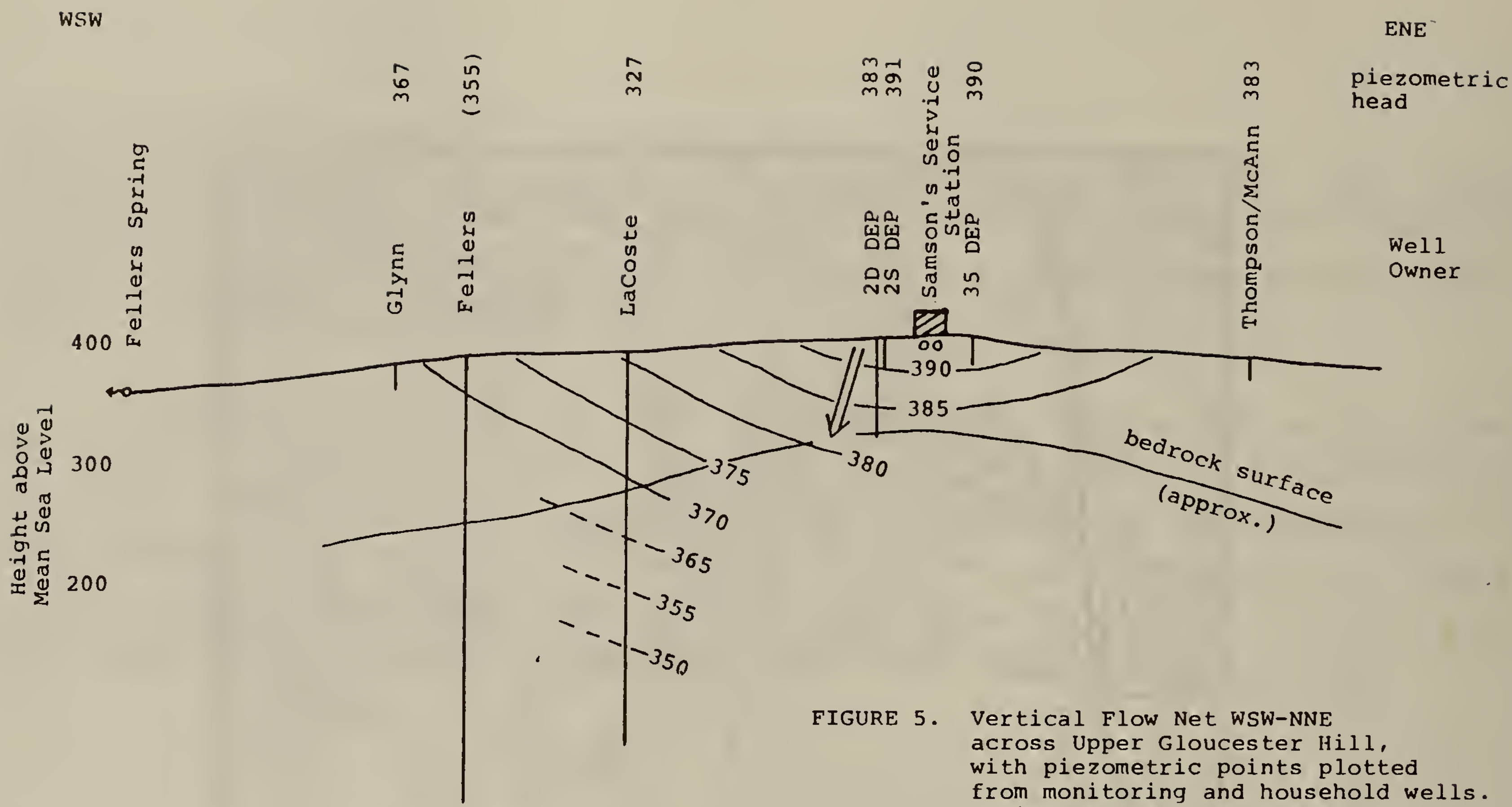


FIGURE 5. Vertical Flow Net WSW-NNE across Upper Gloucester Hill, with piezometric points plotted from monitoring and household wells. Horizontal:vertical scales 1:1.1 Note that the Lacoste water level may be 15 or 20 feet lower than when at equilibrium due to incomplete recovery. Also the Fellers water level was estimated, not measured.

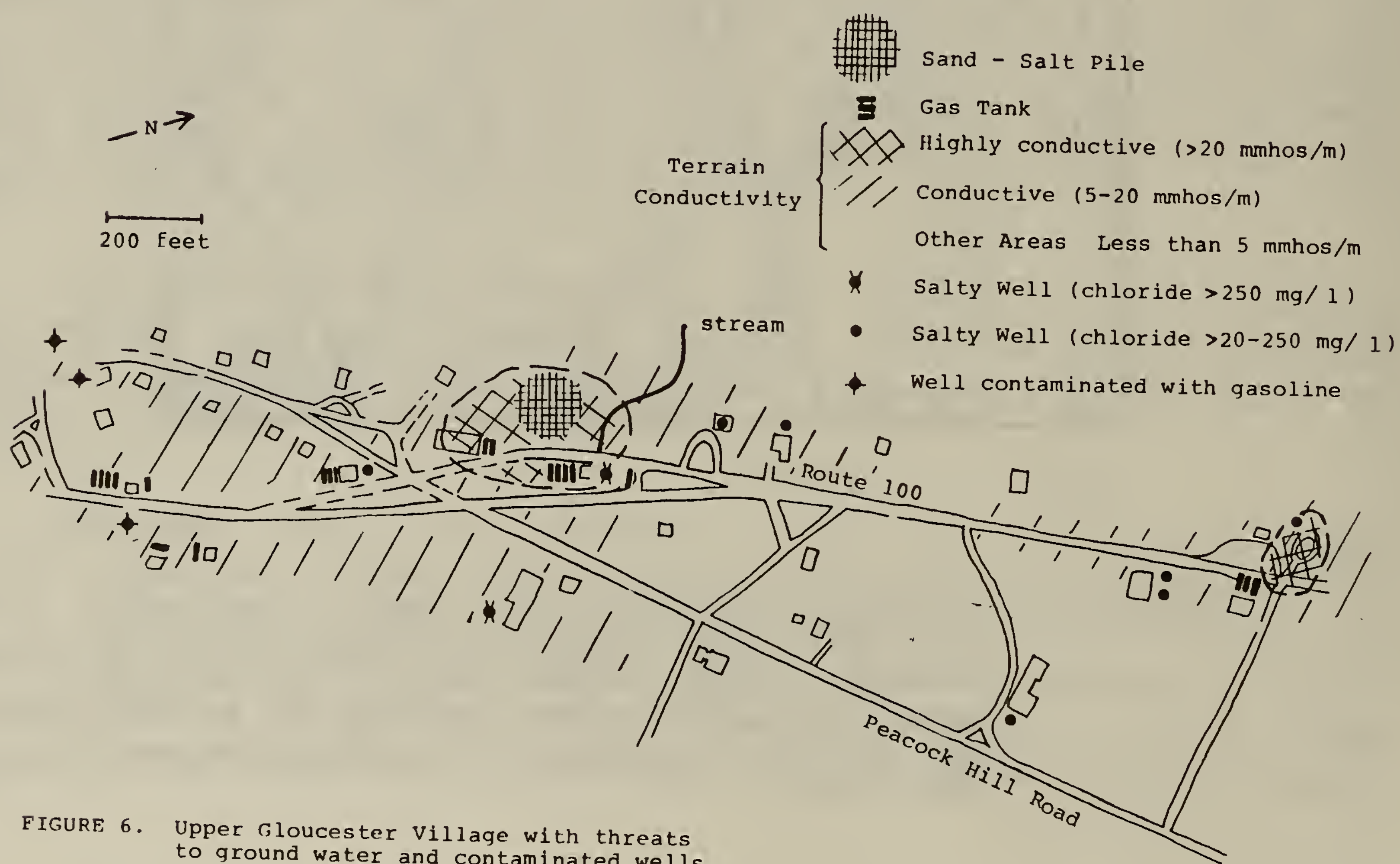


FIGURE 6. Upper Gloucester Village with threats to ground water and contaminated wells.

because the till is so impermeable. Needless to say it takes a long while for water and contaminants to permeate down.

There are two major kinds of threats to ground water in Upper Gloucester. First of all the underground petroleum tanks: we have counted 20 in the village (Fig. 6), and there have been others in the past. Secondly, the sand/salt pile at the town garage.

Around the site of one service station on the hill, there are three household wells (one dug, two drilled) contaminated with gasoline. And near the sand/salt pile there are two drilled wells with chloride exceeding the State drinking water standard of 250 ppm, and several others with elevated levels. Dug wells are better off, as usual in the case of salt. Fig. 6 shows the terrain conductivity contours around the sand/salt pile.

HYDROGEOLOGICAL TOOLS FOR OUR DOG 'N PONY SHOW

Thermometer for measuring ground water temperatures, especially in summer and winter, when they are most different from surface water temperatures.

Ground water flow meter for measuring rate and direction of ground water flow in permeable deposits.

Water level meter for measuring water levels in wells.

Pop level or transit for comparing well elevations to a common datum.

Portable gas chromatograph for sniffing out volatile contaminants in soil and water.

Terrain conductivity meter for detecting electrolytes like salty water.

Voltmeter for measuring the tendency of steel tanks to corrode in soil.

REFERENCES

There aren't any, except in the files of the Maine Department of Environmental Protection. Contaminant hydrogeology is a rapidly evolving science. Even Freeze and Cherry's "Groundwater", published in 1979 makes no mention of chlorinated hydrocarbons as groundwater contaminants. So for further reading on the subject in general, we urge the perusal of current issues of Ground Water, the Ground Water Monitoring Review, and the proceedings of specialist conferences.

ITINERARY

take Maine Turnpike to Gray Exit (#11). Start trip counter at the booth.

0.0 Turn right on Route 202 into Gray, and at the light, turn right south on Route 100.

5.2 At amber flashing light, turn right onto Blackstrap Road, cross over the turnpike, and turn into the Blue Rock pit (#1 on Fig. 1) at **5.7**, where we will park for our walkaround of the West Cumberland site.

return to Route 100 (red light now) at **6.2** Turn left (north). Start counting the number of underground tanks along the way. You can recognise them by the vent pipes with funny little **V** or **T** caps, at the side of service station or other facility buildings.

18.9 Look for big brick Mason's Lodge. This is where we park for our walking tour of Upper Gloucester.

TRIP B-7

SURFICIAL DEPOSITS IN THE LOWER SANDY RIVER VALLEY AND ADJACENT AREAS

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INTRODUCTION

The lower Sandy River Valley in Maine and areas adjacent to it contain excellent examples of surficial deposits representative of styles of deglaciation in Maine, and Pleistocene stratigraphy in New England. Furthermore, several locations are historical, having been described in earlier classic works. The purpose of this trip is to visit some of these locations and provide discussion of them. Weddle has conducted detailed studies at only one of the field stops, but since this summer has worked at a reconnaissance level in the area. Caldwell has been working in the region off and on since the mid-1950's.

PREVIOUS WORK IN THE REGION

The Sandy River Valley has not suffered from an overabundance of detailed surficial geological study. Most work has been of a reconnaissance nature, and until the late 1960's, 7 1/2-minute topographic sheets of the area were not available. Only recently have provisional 7 1/2-minute sheets been produced which nearly cover the entire region. In some areas, however, 15-minute topographic sheets are the only available base maps.

Stone (1899) presented a list of early studies of the geology of Maine: his work contains the first detailed descriptions of deposits in the field trip area. Leavitt and Perkins' (1935) bulletin further described the surficial deposits of Maine. Caldwell (1953) studied the surficial geology in part of the Farmington and Livermore 15-minute quadrangles, and presented more work in later publications (Caldwell, 1959, 1960). The Maine Geological Survey has published open-file reports of surficial maps which cover part of the area (Thompson, 1977; Thompson and Smith, 1977; Smith, 1980). The Surficial Geologic Map of Maine (Thompson and Borns, 1985a) indicates Caldwell and Thompson were responsible for compilation of data in the lower Sandy River Valley.

GEOLOGIC SETTING

The area of interest is shown in Figure 1. It occurs in west-central Maine in Androscoggin, Franklin, Kennebec, and Somerset Counties. The Kennebec River is the major stream in the area, and the Sandy River is tributary to it. Relief in the region varies from 200 to 2000 feet. The bedrock geology of the area is explained on the Bedrock Geologic Map of Maine (Osberg and others, 1985).

Surficial Geology

In general, till and thin drift mantle the upland regions, whereas stratified drift and recent stream alluvium occupy the valleys. Swamps occur where depth to bedrock is shallow, or where the substrate is relatively impermeable.

High-elevation stratified drift deposits, representative of early stages of deglaciation, occur sporadically in the area. One such deposit, comprised of silty fine sand and boulders deposited in a short-lived ice-dammed lake graded to a spillway at 585 feet asl, can be located on the NE corner of the 7 1/2-minute Livermore Falls and NW corner of the 7 1/2-minute Fayette quadrangles.

The recession of active late Wisconsinan ice from southern Maine has been well documented (Thompson, 1982). Features indicative of active ice during deglaciation such as end moraines and ice-shove structures have not been found by the authors in the lower Sandy River area, however end moraines have been mapped by others, and are shown on the state surficial map (Thompson and Borns, 1985a)

Portions of two long esker systems pass through the area, the Chesterville esker system, and the Norridgewock-Smithfield esker-fan-delta complex. These esker systems are part of much larger systems, which occupy the major valleys in Maine. The Norridgewock-Smithfield complex is instructive because along portions of it, glacial marine deltas and fans were deposited, and in some locations mark the approximate position of the ice front and the marine limit during deglaciation. The marine limit varies from approximately 360 feet asl in the southern part of the area, to about 420 feet asl in the north (Thompson and others, 1983).

Throughout much of the area, marine deposits of gray clayey silt and silty clay occur, and are termed the Presumpscot Formation (Bloom, 1960). It often is oxidized to brownish gray, with manganese (?) staining along desiccation surfaces, and generally is massive. Stratification does occur in places, however, and coarser-grained deposits associated with the Presumpscot Formation occur. These sandy deposits are believed to represent late-glacial outwash, graded to a regressing shoreline due to falling sea level. In the upper Kennebec River Valley, this outwash is termed the Embden Formation (Borns and Hagar, 1965), and elsewhere is referred to as the sandy facies of the Presumpscot Formation (Thompson, 1982). Fossil shells collected from the Presumpscot Formation in Norridgewock were some of the first shells from Maine to be radiocarbon dated (11,950 $\pm$ 350 yr BP, W-947, uncorrected whole-shell date, Bloom, 1963).

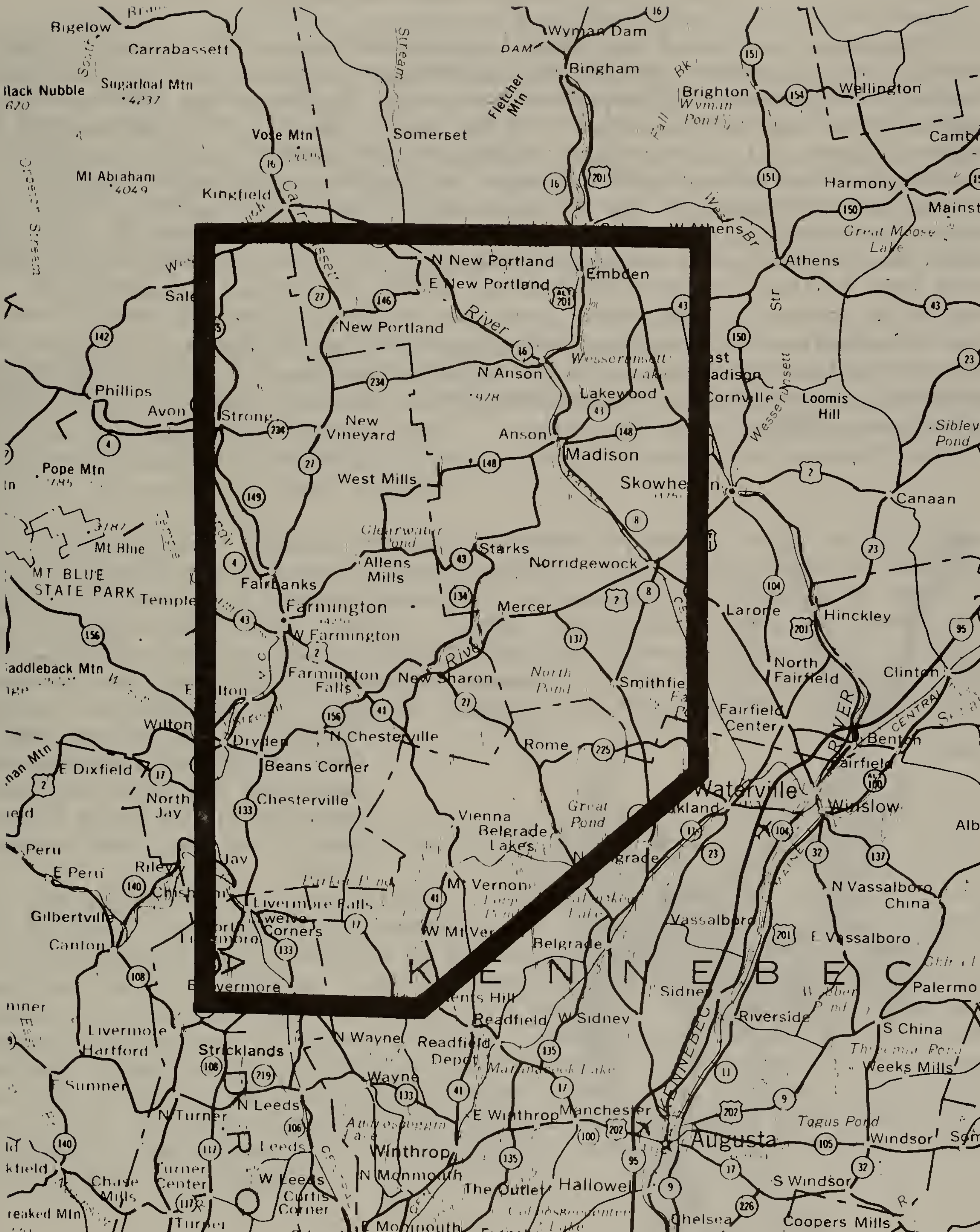


FIGURE 1. Area covered by field trip, lower Sandy River Valley and adjacent areas in Maine. Scale 1:500,000 (refer to field guide for stop locations).

Glacial Lake Farmington

The state surficial map shows the marine limit extent into the lower Sandy River Valley and some of its tributaries. However, Caldwell (1959) and Caldwell and others (1985) suggest that prior to the marine submergence, a glacial lake existed in portions of the valley and its tributaries west of New Sharon (Figure 2). Glacial Lake Farmington was ponded by till deposits at New Sharon, and stratified drift deposits to the south near Twelve Corners. Several possible outlets for the lake were proposed by Caldwell (1959), the highest of which is at approximately 385 feet asl, very close to the marine limit in the field area. Several samples of glaciolacustrine clay collected from test borings, references to large deltas found near the mouths of tributary streams to the Sandy river, one exposure along the Sandy River, and the spillways are the proxy evidence by which the existence of Glacial Lake Farmington is established. The lake ended when the till dam at New Sharon was breached, and the valley became inundated by the sea (Caldwell, 1959; Caldwell and others, 1985).

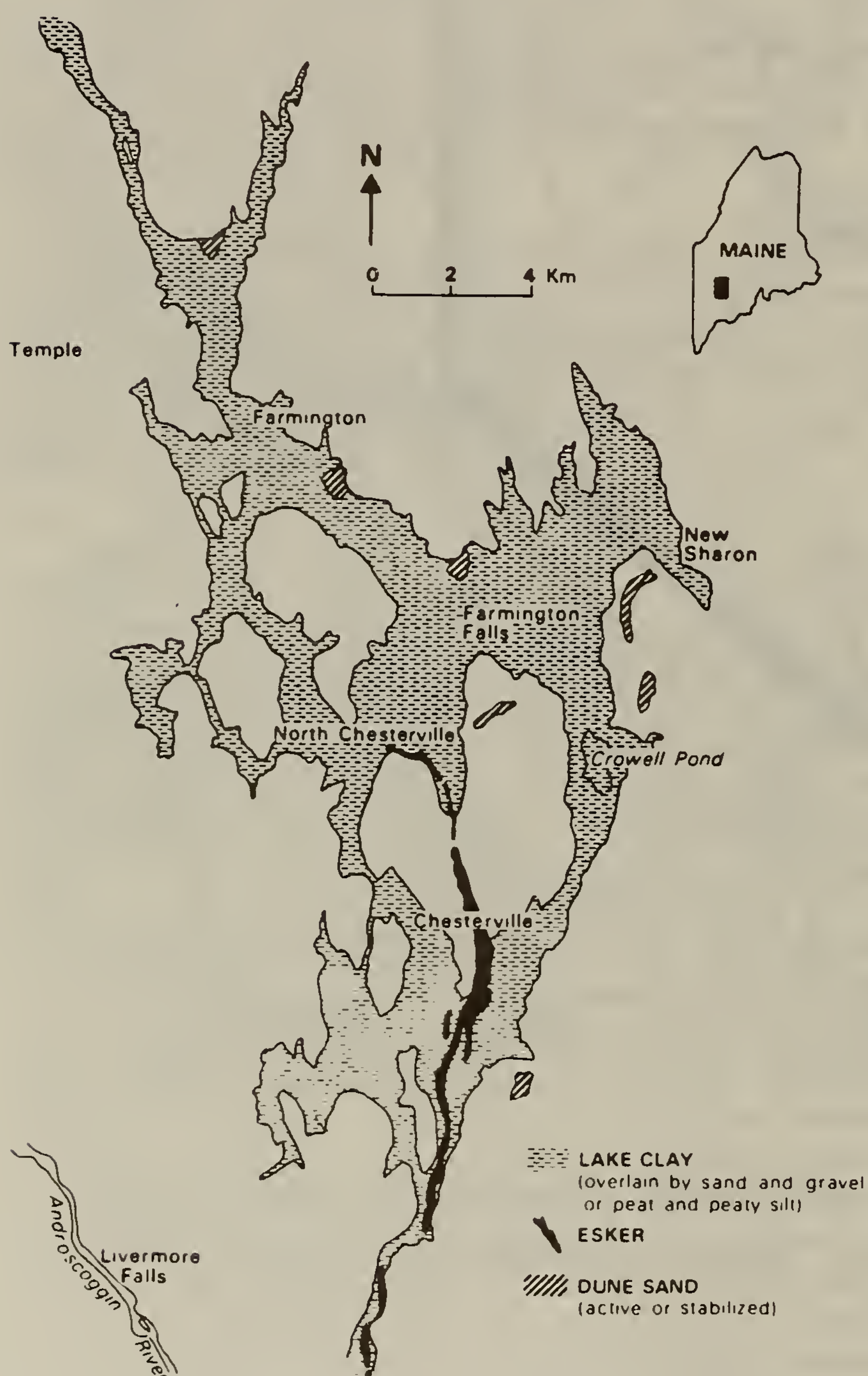


FIGURE 2. Map showing the distribution of lake-bottom sediments and other surficial deposits in the Sandy River basin (Caldwell and others, 1985).

The Little Norridgewock Stream (Fayette, Farmington Falls, Livermore Falls, Wilton 7 1/2-minute quadrangles) and its tributaries are northerly flowing streams. If these valleys were free from marine water invasion for a brief period, conditions may have been amenable for existence of a freshwater lake. If this was the case, then evidence for an ice-dammed Glacial Lake Farmington should be found in the valley. Field reconnaissance (Weddle) to date indicates some ice-marginal positions occur in these valleys, and that local glacial-lake ponding occurred, however there appears to be little direct evidence for a large lake the size of Glacial Lake Farmington. Foraminifera have been identified from sediments within the Glacial Lake Farmington basin, suggesting that marine conditions were predominant in the area (H. W. Borns, Jr., pers. comm., 1986).

Ice-contact stratified drift deposits south of Chesterville occur at about 430 feet asl. Southward in this constricted valley, kettled stratified drift occurs at about 400 feet asl just south of Twelve Corners, and to about 380 feet asl slightly farther south. A delta (?) one mile east of Livermore Falls has a surface elevation of 383 feet asl, however no topset/foreset contact is exposed in this feature.

Approximately twenty-five miles farther northeast, north of Smithfield, a glacial marine delta topset/foreset contact is shown on the state surficial map at 375 feet, and another contact 15 miles east in Belgrade is at 359 feet asl (Thompson and Borns, 1985a). In the area south of Twelve Corners, a glacial marine limit between 360-370 feet asl seems likely. Most of the glaciofluvial and glaciolacustrine deposits in the Little Norridgewock Stream valley, at least as far north as Chesterville, are probably graded to this level.

Till Stratigraphy at New Sharon

The gorge of the Sandy River, at New Sharon, west-central Maine, is one of the best exposures of Quaternary deposits in New England. It is also the only recorded surface locality in New England where organic material is reported to occur between older and younger aged tills (Shafer and Hartshorn, 1965; Mickelson and others, 1983). The original location of the organic material is covered by river alluvium. Re-excavation of the site by backhoe in 1985 proved ineffectual in documenting the earlier described stratigraphy because of ground water seepage into the trench. Part of this field trip will be to view the stratigraphy at New Sharon, and provide an opportunity to discuss its relationship to stratigraphies outlined for southeastern Quebec and southern New England.

Previous Work at New Sharon

Caldwell (1959, 1960) presented the earliest work on New Sharon. He interpreted the deposits there to be the products of two episodes of glaciation. These glacial events were separated by a non-glacial period, during which a soil developed on an organic-bearing silt, overlying till deposited from the first glaciation (Caldwell, 1959). Fragments of wood

collected from the silt were dated at greater than 38,000 yr BP (W-910), and pollen analyses of the organic-bearing material indicate that a colder climate than present existed. An interstadial age, middle Wisconsinan(?), was assigned to the soil (Caldwell, 1960). On the basis of the radiocarbon date, pollen assemblage, and stratigraphy, the buried soil was tentatively correlated by Caldwell (1960) with non-glacial deposits in Canada, the St. Pierre Sediments, which were described by Terasmae (1958, 1960). Caldwell (1959) attributed both glacial events to the Wisconsinan glacial stage; the deposits under the organic horizon were assigned an early Wisconsinan age, and the units over the organic horizon a late Wisconsinan age. Caldwell and Pratt (1983) later suggested that till of middle Wisconsinan age may also be present at the New Sharon section.

Borns and Calkin (1977) collected a continuous core of glacial sediment from a test boring at New Sharon, however, the boring did not penetrate the organic-bearing silt. They had the wood previously collected from New Sharon redated, with a resultant age of greater than 52,000 yr BP (Y-2683) determined. With interpretation of the core record, previous work by Caldwell (1959, 1960), and the new radiocarbon date, Borns and Calkin (1977) suggested that the Quaternary stratigraphy at New Sharon is equivalent in age to the stratigraphy in southeastern Quebec and the central St. Lawrence Lowlands, as determined by McDonald and Shilts (1971), Gadd (1971), and Gadd and others (1972). This correlation is significant because the stratigraphy in southeastern Quebec is attributed to three glacial episodes, separated by non-glacial events.

Age of Tills and Regional Considerations

Southeastern Quebec. In the Appalachian highland region of southeastern Quebec, Wisconsinan glacial deposits are interpreted in a threefold stratigraphy, representing early(?), middle, and late Wisconsinan events (LaSalle, 1984). McDonald and Shilts (1971) describe an early Wisconsinan(?) till, the Johnville till, as the oldest till in the region. Overlying this unit is the Massawippi Formation, a partly fluvial and partly lacustrine unit, from which organic material has been collected and dated at greater than 54,000 yr BP (Y-1683).

The next youngest till is the Chaudiere till of middle Wisconsinan age. This till is distinctive lithologically, chemically, and mineralogically from other tills in the region, and was deposited by ice flowing originally to the southwest, and later by ice flowing southeasterly (McDonald and Shilts, 1971). Shilts (1981) postulates a Maritime source for the Chaudiere event, and attributes the shift in flow to later arrival of Laurentide ice, which merged with or displaced Maritime ice during middle Wisconsinan time.

Overlying this till is the Gayhurst Formation, a glaciolacustrine unit, dated at greater than 20,000 yr BP (GSC-1137). The youngest till in the region is the late Wisconsinan Lennoxville till, which overlies the Gayhurst Formation. The Lennoxville till is separated into two members; a brown, oxidized, loose sandy upper member, and a gray nonoxidized, jointed, compact clayey lower member (Shilts, 1978; Chauvin, 1979).

The criteria which determine this stratigraphy are based on mineralogical and lithological variations, which consistently occur upsection and are noted regionally, and on radiocarbon dates from the Massawippi and Gayhurst units (McDonald and Shilts, 1971; Shilts, 1979, 1981; LaSalle, 1984).

Studies in southeastern Quebec by Shilts (1978, 1981), and by Chauvin (1979), have described sections stratigraphically similar to the section at New Sharon. Further work in southeastern Quebec by Parent and others (1984), have incorporated site-specific studies of several Quaternary sections into a regional context, summarizing Wisconsinan events in the area. These workers have not attempted any correlation of the Quebec stratigraphy with New Sharon. In a summary of the Quaternary stratigraphy of Quebec, LaSalle (1984) suggests more work and better sections in New England are needed to clarify any correlations.

Southern New England. In southern New England, Quaternary stratigraphy is interpreted by most workers there to represent only two glaciations (Currier, 1941; Moss, 1943; White, 1947; Schafer and Hartshorn, 1965; Pessl and Schafer, 1968; Pessl and Koteff, 1970; Newton, 1978, 1979; Koteff and Pessl, 1985).

Tills associated with the two glaciations have long been referred to as "old" or "lower" till, and "young" or "upper" till. Some of the field characteristics which distinguish these units include stratigraphic position (usually upper over lower where found in contact), compaction (generally the lower till is more compact that the upper till), texture (upper till is commonly sandier than the lower till), color (gray upper over olive gray or brown lower), structural relationships at the contact between the two units, and depth of weathering of the units (upper till less deeply weathered than lower till).

The age of the two glacial events in southern New England is thought to be early Wisconsinan for the first, and late Wisconsinan for the second (Schafer and Hartshorn, 1965). Deposition of sediments associated with these two events was interrupted by an interval during middle Wisconsinan time, when weathering of the early Wisconsinan deposits occurred (Schafer and Hartshorn, 1965; Newton, 1978). This weathering is represented by an oxidized zone of 10 m or more on the surface of the lower till. Newton (1978) described the mineralogical characteristics of the oxidized lower till, including detailed analyses of the clay minerals associated with it. He concluded that the oxidized zone of the lower till represents weathering, probably associated with soil formation during middle Wisconsinan time. He also proposed a lithostratigraphy for the tills of southern New England, formalizing the terminology from lower and upper tills to Thomaston and Bakersville tills, respectively, with the type localities in Connecticut.

The stratigraphy at New Sharon described by Caldwell (1959, 1960), includes mention of a weathered appearance to the surface of the organic-bearing silt. Koteff and Pessl (1985) correlate the stratigraphy at New Sharon described by Caldwell (1959, 1960), with the stratigraphy exposed along Nash Stream, New Hampshire, their reference section for the two-till stratigraphy of southern New England (Pessl and Koteff, 1970; Koteff and Pessl, 1985). They do not correlate middle Wisconsinan units of southeastern Quebec with any deposits in southern New England, and suggest that Borns and Calkins's (1977) interpretation of New Sharon is better correlated with Nash Stream than with the Quebec stratigraphy.

Maine. In Maine, a two-till stratigraphy was described by the early workers, although agreement on whether the tills represent multiple glaciation was never established (Holmes and Hitchcock, 1861; Stone, 1899; Clapp, 1906, 1908; Leavitt and Perkins, 1935). New Sharon was known by Leavitt and Perkins to be an organic-bearing locality (pers. comm., H. W. Borns, Jr., 1986), however they do not mention it in their Bulletin No. 30 (Leavitt and Perkins, 1935). After Caldwell's discovery of the wood at New Sharon (1959), the two-till problem, at least in west-central Maine, became unequivocal (Schafer and Hartshorn, 1965). Detailed studies were not done at New Sharon after Caldwell's initial work (1959, 1960), except for Borns and Calkin (1977). Since then, and until the early 1980's, studies on till stratigraphy were focused elsewhere in the state (summarized in Kite and others, 1986).

Till stratigraphy in southern Maine has been reviewed by Thompson and Borns (1985b). They propose a stratigraphy in southwestern Maine correlative with Nash Stream, after Koteff and Pessl (1985). New Sharon is discussed in their article, but no correlation of it with any other stratigraphy is presented. However, because of the compacted appearance of wood specimens from New Sharon examined by F. Hyland (pers. comm., H. W. Borns, Jr., 1986), Thompson and Borns (1985b) state that the wood may have been killed by overriding ice, and that much of the section at New Sharon could be pre-late Wisconsinan in age.

The wood which the present authors have found in 1959 and 1985, has been in fragments, and photographs of wood in till taken at New Sharon in 1959 do not show trees in growth position, but fragments of wood. Other organic material (peat) found in 1985, occurs in sandy gravel underlying the wood-bearing silty unit. It appears that these organic materials were transported and deposited, either fluvially or in standing water. Any indication of how the wood died is not obvious in hand specimens. Caldwell (pers. comm., 1985) noted charcoal fragments in the silty unit, but no large, charred wood fragments have been uncovered.

Dr. Richard Jagels (Forest Biology, University of Maine, Orono) examined the wood found in 1985, however the specimens were too desiccated to be identified. He did not note any indication from these samples as to how the wood may have died, and believes the compaction of specimens is more likely due to compaction by overlying sediments (pers. comm., R. Jagels, 1986).

We do not believe the age of the wood constrains the age of the overlying till to pre-late Wisconsinan. One of us (Caldwell) believes there may be middle and early Wisconsinan units at New Sharon, however we both agree there is late Wisconsinan till present.

A test boring in Newport, Maine, records a subsurface stratigraphy of two tills separated by an organic-bearing fine-grained unit (pers. comm., H. W. Borns, Jr., 1986). The upper till in the test boring changes color with depth, from olive-brown to gray. This stratigraphy is similar to that initially described by Caldwell (1959) at New Sharon, approximately 60 km to the west.

It is not the intent of the authors to propose correlation between deposits described from the test-boring records with the surface exposures at New Sharon. However, it is obvious that with subsurface data, a better

understanding of the Quaternary history of Maine can be developed. Geotechnical and hydrogeological firms are commonly providing better documentation of test-boring records. Geologists with these firms should interact with academic and governmental geologists when valuable subsurface data becomes available.

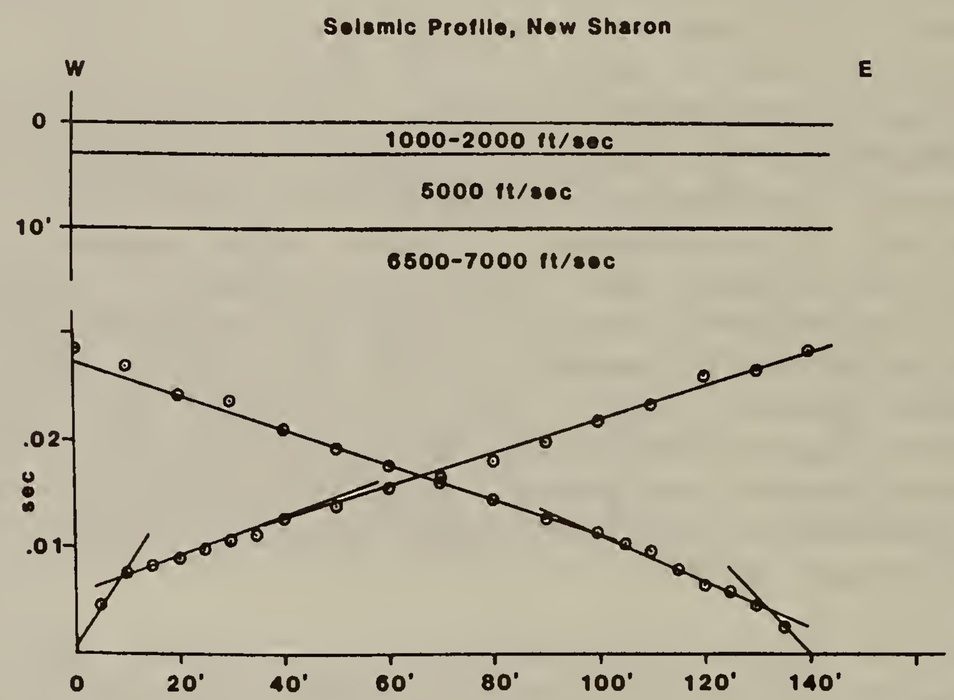
Recent Work

The obvious conflict between regional stratigraphic correlation of Quaternary deposits in New England and Quebec has spurred recent work in Maine, undertaken in an effort to clarify the stratigraphy and sedimentology at New Sharon and its relationship to those other regions (Caldwell and Pratt, 1983; Caldwell and Weddle, 1983; Weddle and Caldwell, 1984; Weddle, 1985). Most recently, the organic-bearing horizon was re-excavated to examine the stratigraphic relationship between it and the glacial deposits it was reported to occur between (Weddle, 1986). This horizon has been described by others as a brown silt, an organic silt layer, buried non-glacial sediments, a buried soil, a subaerial weathering profile containing organic debris, and the New Sharon soil (Caldwell, 1959, 1960; Borns and Calkin, 1977; Mickelson and others, 1983; Hanson, 1984).

The excavated sequence includes 0.5 m of glacial deposits over 1.0 m of organic-bearing silt, which overlies 1.5 m of sandy gravel. The sandy gravel is nonoxidized below the silt, however 1.0 m below this contact it gradually changes color from gray (2.5YR6/6) to reddish yellow (5YR6/8 to 7.5YR6/8). The organic-bearing unit displays no obvious weathering profile or soil horizon, and consists of complexly deformed, thinly laminated, alternating very dark to dark grayish brown (2.5Y3/2-4/2), and dark gray to olive gray (5Y4/1-4/2) fine sandy silt and silty fine sand layers. The contact between the different colored layers is sharp, and appears to be lithologically controlled. No till was encountered under either the silt or the gravel, however, excavation had to be ceased due to ground water seepage into the trench. Seismic refraction lines run parallel to the trench indicate velocities consistent with the interpretation that till occurs under the organic-bearing horizon (Figure 3). Pollen analyses performed on samples collected from the organic-bearing unit are consistent with the interpretation that it represents an interstadial deposit, rather than an interglacial deposit (pine-spruce-alder pollen floras, very little hardwood; R. Nelson, pers. comm., 1985.

The numerous references to the weathered appearance of the surface of the organic-bearing unit do not concur with the appearance of this unit during excavation in 1985. As previously stated, no obvious soil horizon is present from the surface downward. Eight samples from the olive-brown and gray layers in the fine-grained unit were analyzed by x-ray diffraction for evidence of clay-mineral alteration indicative of subaerial weathering. All samples run indicate the same pattern. The results, partially shown in Figure 4, are compared with x-ray diffraction patterns from the oxidized lower till of southern New England (Newton, 1978). There is no alteration of any of the minerals from the organic-bearing unit at New Sharon, as indicated by the various tests run on the samples (air dried, ethylene-glycol bath, K-saturation, various heat treatments). More importantly, there is no

FIGURE 3. Seismic profile, New Sharon, Maine; 5000 ft/sec section represents organic-bearing horizon (silt and sandy gravel), 6500-7000 ft/sec section most likely reflects underlying till.



alteration of chlorite in the samples from New Sharon, unlike the samples of oxidized lower till from southern New England, which show decreasing alteration of chlorite with depth (Newton, 1978).

It appears that the organic-bearing unit at New Sharon has never been subjected to weathering comparable to that proposed for the lower till of southern New England. The organic-bearing unit at New Sharon cannot be considered a soil because neither soil nor weathering profiles occur there. Hence, if correlation of New Sharon with sites elsewhere in New England or Canada is based on a weathering horizon the correlation is not warranted.

It is apparent that further work is necessary to document the existence and nature of an older, underlying till at New Sharon. Although reevaluation of previous interpretations is warranted for the stratigraphy at New Sharon, field work to date suggests the complete section exposed along the river appears to be the product of late Wisconsinan events. Regional interpretations based on the former stratigraphic interpretations, and the origin of the organic-bearing horizon also need to be reevaluated.

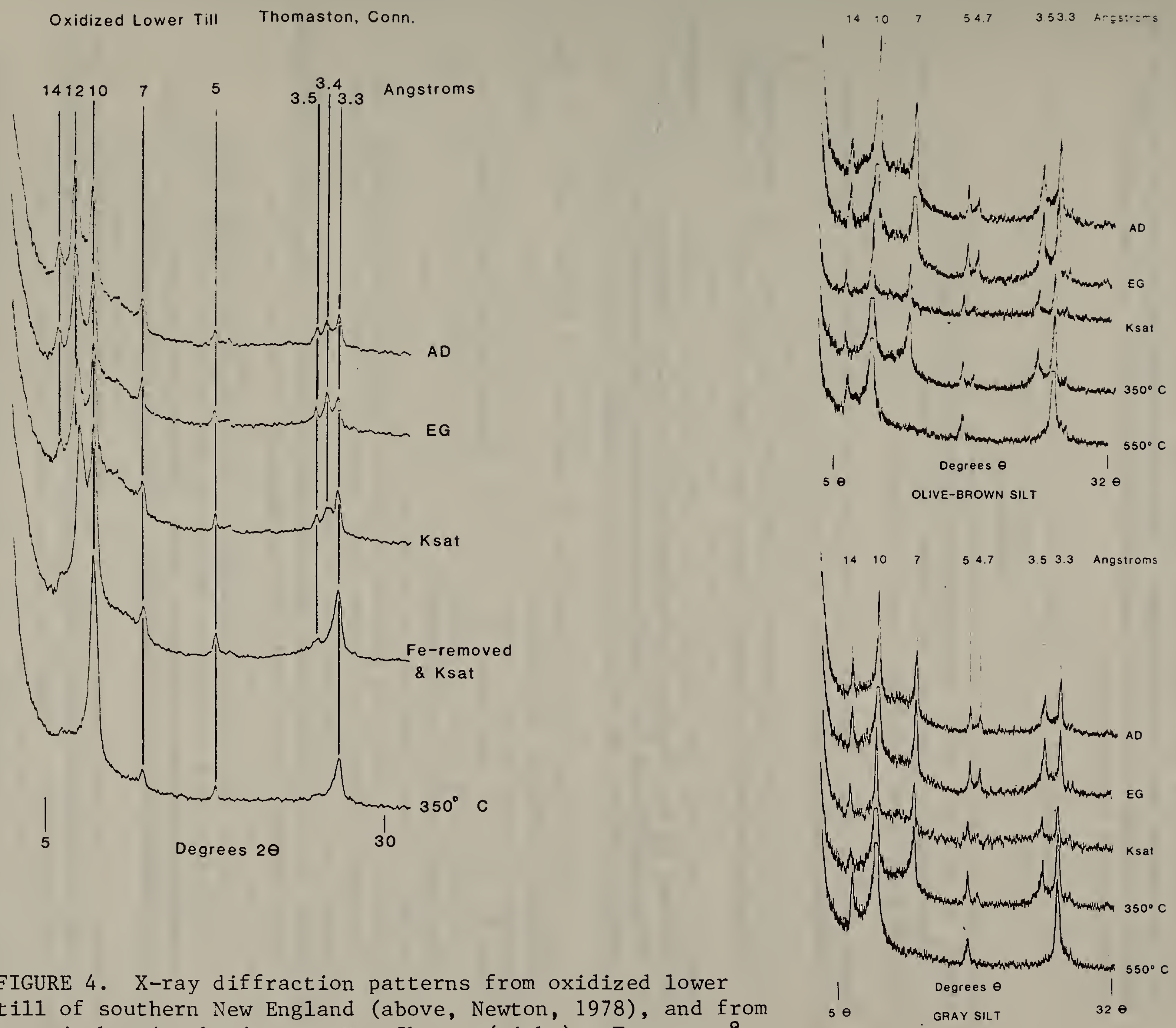


FIGURE 4. X-ray diffraction patterns from oxidized lower till of southern New England (above, Newton, 1978), and from organic-bearing horizon at New Sharon (right). Fourteen Å peak on sample from southern New England represents weathered chlorite; note lack of alteration of 14 Å peak on New Sharon sample patterns.

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ITINERARY

Assembly point is at Norridgewock Center at intersection of Routes 2 and 139. From Interstate 95, take Exit 34 north on Route 104 to Route 139 and into Norridgewock. Assembly time is at 9:30 A.M. Topographic maps include Anson, Norridgewock (15-minute); New Sharon, Farmington Falls, Fayette, Wilton, Livermore Falls (7 1/2-minute); provisional 7 1/2-minute sheets are available for Norridgewock 15-minute sheet.

Mileage

0.0 Junction of Routes 139 and 2--Follow Route 2 west through Norridgewock Center.

0.4 Turn right (north) on Route 8/201A.

0.6 Cross Kennebec River.

6.2 Gravel pits; Madison esker-fan-delta-complex.

6.4 Madison-Norridgewock town line.

8.1 Junction of Routes 8/201A with 43/148; turn left (west) at lights, following 8/201A.

8.5 Cross Kennebec River from Madison into Anson, and continue north on 8/201A (right turn over bridge) along flood plain and terraces of river.

13.2 Cross Carrabassett River, enter North Anson--continue north 8/201A.

13.3 Turn right at North Anson Community Church (Madison Street); drive along river terrace.

13.4 Descend to flood plain.

13.8 Paved road ends.

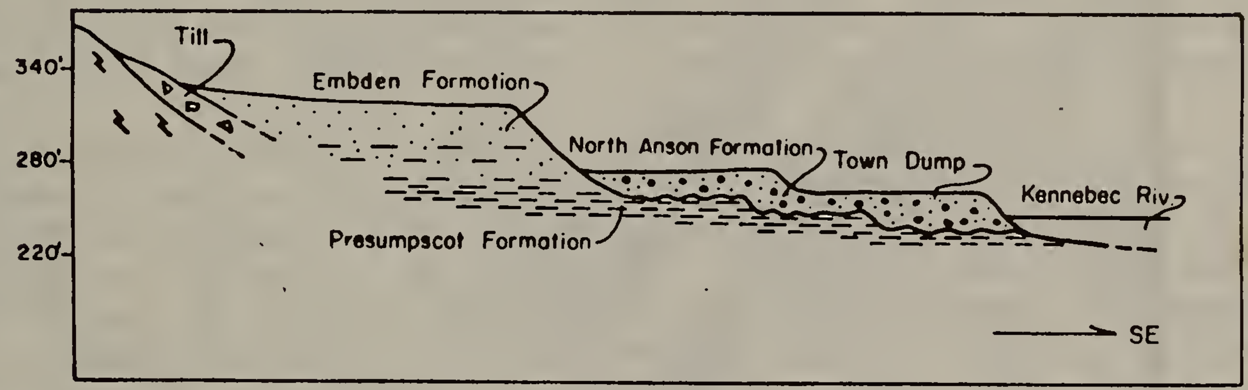
13.9 Cross under power lines; stream terrace to left.

14.6 Gravel pit in lower terrace deposits (fluvial sand and gravel).

15.0 STOP 1. Anson-Embden Landfill (Anson 15-minute sheet); coarse subaqueous outwash interbedded with and overlain by fine-grained silty sand, silt, and clay layers, in turn unconformably overlain by coarse grained fluvial deposits.

This section has been described by Borns and Hagar (1965) and the relationships between various deposits and terrace levels is schematically represented in Figure 5. Following their terminology, the exposure represents deposition of ice-proximal subaqueous outwash followed by deposition of Presumpscot Formation. This unit grades into the sandy Embden Formation, which represents the regressive facies of the marine inundation. The Embden Formation has been truncated by the overlying North Anson Formation, representative of late glacial outwash attributed to localized events in the Kennebec River drainage (Attig, 1975).

FIGURE 5. STOP 1. Schematic cross section showing the stratigraphy on the west side of the Kennebec River valley at North Anson, Maine (Borns and Hagar, 1965)



Sedimentary features present at this locality include a general coarsening-upward in the fine-grained facies, syndepositional drag folds, load structures, dropstones, clay intraclasts, and cut-and-filled channels.

Return to vehicles; continue north on Madison Road.

15.4 Stop sign; Junction with Routes 8/201A (Arnold Highway), flood plain of Kennebec River to right (east); turn left and ascend higher terrace level (sandy gravels of Embden Formation).

17.2 Junction Routes 8/201A with Route 16, North Anson center; retrace route back to Norridgewock following Routes 8/201A.

30.0 Junction Route 2 Norridgewock; take sharp right (Winding Hill Road); drive along terrace.

30.3 Descend to Kennebec River flood plain.

30.9 STOP 2 (Figure 6). Park on left or right but BEWARE OF SOFT SHOULDERS - DO NOT GET STUCK!!! Norridgewock esker-fan-delta complex (Norridgewock 7 1/2-minute provisional sheet). This pit is historical because as previously noted it is where in Maine some of the first fossil marine shells were collected from and dated (Bloom, 1963). *Mytilus edulis* shells, whole and in fragments are presently exposed in the Presumpscot Formation excavated just south of the parking area.

This large excavation exposes a spectrum of deposits varying from very coarse-grained subaqueous outwash to fine-grained marine silt and clay. In the center of the pit, exposed just under 300 feet asl is a pebble lag overlying the Presumpscot Formation. Over the pebble lag a few feet of medium- to fine-grained sand with low-angle, planar and trough cross-bedding occurs. This deposit may represent the regressive shoreline of the falling sea level after the maxiumum marine submergence. NOTE: This feature is not easily accessible and requires

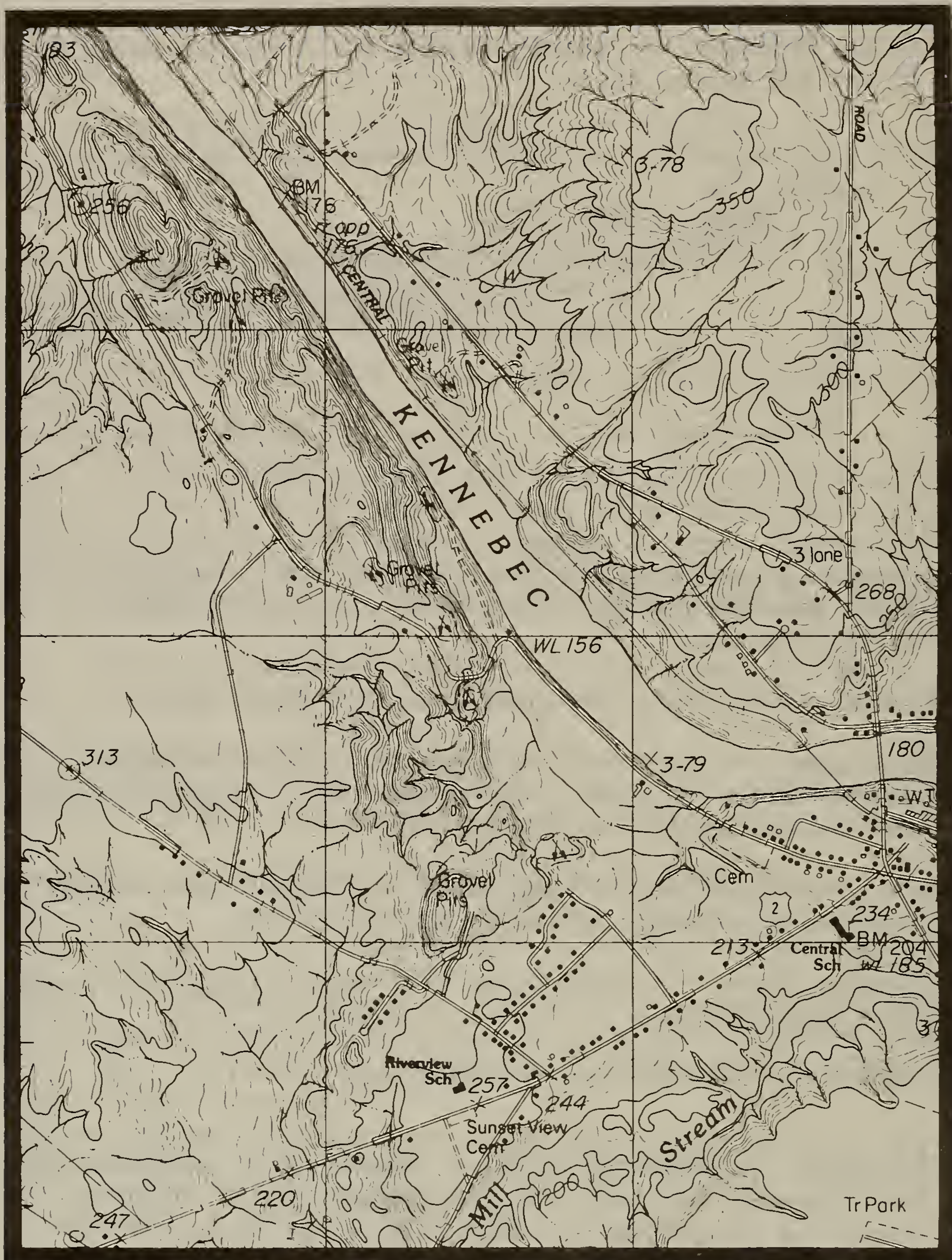


FIGURE 6. STOP 2. Norridgewock Esker-Fan-Delta Complex. Norridgewock 7½' Provisional sheet, scale 1:24,000.

more time and stamina to view than many may want to exert--please do not wander far from the group or remain away too long if you do want to see this feature.

The relationships represented here are similar to those seen at Stop 1, however they occur on a grander scale. Ice-contact collapse features are present, as well as sedimentary structures such as climbing ripple drift, and drag folds; grain-flow and debris-flow deposits occur.

The cumbersome terminology applied to this feature (Norridgewock esker-fan-delta complex) is due in part to the town name applied to the pit, and the morphology of the deposits associated with it. Much of the topography appears to be controlled by collapsed subaqueous outwash, ice-blocks, kettles and planation by the falling sea. One portion of the complex reaches an elevation of just over 410 feet asl (Figure 6), and although no topset/foreset contact is exposed here, the topography and elevation of this part of the complex suggest this may be the only truly deltaic portion of it. This part of the complex appears to be esker fed as well.

Return to vehicles; continue up Winding Hill Road.

31.5 Turn left at dirt road; "deltaic" portion of complex to right.

32.3 Stop sign, turn left.

33.1 Stop sign, turn right (Route 2 west) and take immediate left fork (Wilder Hill Road); travel along 257 feet asl surface short distance (regressive sea surface).

33.7 Presumpscot Formation beneath sand plain to right.

36.6 Turn left, Sand Hill Road.

38.1 STOP 3. Little Pond Delta (Norridgewock 7 1/2-minute provisional sheet). LUNCH STOP. Excellent example of an esker fed delta. Much of the delta is collapsed, however a topset/foreset contact measured at 375 feet asl is shown on the new State Surfical Map. The flat-topped, uncollapsed part of the delta we will have lunch on is at approximately 390-400 feet asl; the 375 feet asl contact thus appears reasonable as an indicator of the marine limit. This will be important to the discussion of Glacial Lake Farmington later in the trip.

Return to vehicles; reverse direction and retrace route on Sand Hill Road, continue past junction with Winding Hill Road.

39.7 T-junction (stop), turn right onto Oak Hill Road.

41.4 Stop sign; turn left onto Route 2 west.

42.6 Junction Routes 2 and 137.

47.0 New Sharon/Mercer town line.

49.2 Turn sharply right immediately before Junction of Route 27.

49.8 Turn left at steel gate: PERMISSION MUST BE REQUESTED FROM LANDOWNER BEFORE ENTERING!

50.0 STOP 4. Sandy River Section (Mercer 7 1/2-minute provisional sheet). This stop is the first of two (or time permitting possibly three) where downcutting by the Sandy River has provided some of the best exposures of Pleistocene deposits in west-central Maine.
Much discussion regarding the use of the term diamicton rather than till has been generated in the past few years. To help keep this argument in perspective, it appears that at the turn of the 19th century discussion about the use of the term till was as lively then as it is now, as described in Stone (1899):

"Resting upon the glaciated rock ... is the till. It is an endless study....

The names given to the till in Maine deserve notice. A very common name for the Formation is 'hardpan' .... Another common name is 'pin gravel', ... 'hard, rocky land', ... 'rocky, upland soil', ... 'hardwood soil', also 'orchard land' .... 'Gravelly loam' almost always means till, but sometimes it means a thin stratum of marine clay overlying and partially mixed with true water-assorted and rounded gravel. Many know the formation as the 'bowlder clay'. To apply the terms 'gravel' or 'clay' to the till is a fruitful source of confusion, causing the till to be confounded with water-washed gravel on the one side and with sedimentary clay containing boulders on the other. The term 'bowlder clay' may still have its uses, to describe certain disputed formations, but in New England it ought to be replaced by the word 'till'. The word is short, convenient, and implies no theory either as to the composition or the origin of the deposit. The till constitutes what was known to the older geologists as the 'drift' or 'unmodified drift'."

One must wonder what Stone would have thought of the term "diamicton"? Undoubtedly, throughout this field trip, the terms "till" and "diamicton" will be used interchangeably during discussion, however for this written guide the term "till" will be used to describe material which is a heterogeneous mixture of clay, silt, sand, and boulders deposited directly by glacial ice, which may subsequently have been remobilized but not undergone significant sorting.

Walk along road toward river. On left is an exposure of silt, silty sand and sand interbedded with sandy flowtill, overlain by gravelly sand which grades into sandy till. Overlying this sequence is the Presumpscot Formation. The lower fine-grained deposits are deformed and contain numerous dewatering structures and cross-cutting stringers of sandy till. Most of this part of the section appears to have been deposited subaqueously.

Continue down road and turn left to large exposure along river. Here, deformed silty sand and interbedded compact till is overlain by

deformed laminated silt and fine sand. This is overlain by gravelly sand, also containing dewatering structures, which grades upward into sandy till containing striated clasts. Overlying this sequence is the Presumpscot Formation.

The exposure is interpreted as proglacial subaqueous sediments deposited in front of advancing ice, subsequently overlain by till deposited by that ice. The Sandy River is flowing northeasterly here, and joins the Kennebec River just south of Madison. It appears that late Wisconsinan ice in the Kennebec Valley dammed the junction of these rivers prior to occupying the lower Sandy River Valley, thus creating conditions for a proglacial lake in the Sandy River Valley.

Return to vehicles, and back to paved road. Turn right at gate.

50.6 Stop sign, turn right on Route 2 west.

51.2 STOP 5. (New Sharon 7 1/2-minute sheet) Turn right onto shoulder at Farmington and Mt. Vernon road signs; park along dirt road and walk through fields (if possible we may drive in). PERMISSION MUST BE REQUESTED FROM LANDOWNER BEFORE ENTERING! Walk along road to culvert and turn left down slope to river.

This long exposure shows the same relationship as described at the previous stop, that is, proglacial subaqueous deposits overlain by late Wisconsinan till. Here, the lower units are comprised of thinly laminated clay, silt, and sand layers interbedded with thin to thick till layers. These units appear to be similar to the Gayhurst Formation of Shilts (1981).

Farther downstream beneath these laminated sediments is a concretion-bearing clayey silt containing striated boulders. Striations on stones and long axes of clasts in this unit do not show a preferred orientation, and it is interpreted as a facies of the laminated unit, deposited subaqueously. Shilts (1981) has described a similar appearing unit in Quebec, the Drolet Lentil, as a basal till drived from underlying clay-rich lacustrine units. Caldwell (1959) decribed this unit as a "boulder clay" (note spelling--somehow he dropped the "w"), and correlated it with the lower till under the organic zone he described from a section on the opposite side of the river. X-ray diffraction analysis of clay minerals from the "boulder clay" and the overlying laminated deposits have no significant difference in patterns (Weddle, 1985).

If time permits, and there is adequate exposure, we may visit the site (on opposite bank) where Caldwell (1959) found the wood.

Return to vehicles, and continue west on Route 2.

57.9 Entrance to Farmington landfill on right, enter through gate and continue on.

58.0 STOP 6. Farmington Landfill. (New Sharon 7 1/2-minute sheet) Park along road and walk to excavated areas. Till and stratified drift contact approximately 0.15 miles to west along dirt road. Elevation is approximately 420 feet asl. Return back toward vehicles. Excavation in fluvial sand to right. Concave- and convex-up surfaces between sedimentary units reflect bar and channel morphology of stream. Clasts and layers of reworked till in the sand observed at the till/stratified drift contact suggest some contribution to the stream by local topography. However, there is a lack of large clasts in the sand exposed at the large excavation, which one would expect to find if nearby till was the source of the sand. The sand is associated with the regressive sequence formed during falling sea level.

Return to vehicles; and reverse direction and drive out to Route 2; turn right (east).

59.1 Junction of Routes 2 and 41/156; turn onto Route 41/156 south to Farmington Falls.

59.5 Cross Sandy River at Farmington Falls.

59.6 Turn left off route 41/156 after crossing bridge.

60.4 Meander scar and infilled oxbow lake on left (Sandy River golf course, Par 3).

70.0 STOP 7. Cut bank along Sandy River (Farmington Falls 7 1/2-minute sheet). Two good exposures in the cut bank show rythmically bedded silt and clay at river level grading upward into plane bedded and climbing-ripple drift sand; the angle of climb is pronounced by iron-oxide staining of the sand. Caldwell (1959, 1960) interpreted these deposits as evidence for glacial Lake Farmington. The elevation at these sections is below 340 feet asl, well below the level of the marine limit. The timing of the deposition of these sediments is not clear. They may have been deposited prior to the marine incursion into a locally-ponded freshwater glacial lake. However, it is more likely they are associated with the regressive deposits observed at Stop 6.

Return to vehicles; reverse direction and return to Route 41/156.

71.4 Junction of Route 41 and 156, continue straight across to Route 156.

71.8 Left turn off Route 156, south to Chesterville.

Coarse gravel and gravelly sand, and coarse- to fine-grained sand laid down by streams associated with nearby ice occur in gravel pits in this area. Leavitt and Perkins (1935) state that this region was flooded by the sea and contains deposits of marine clay and sand. However, no marine sediments overlie the coarser fluvial deposits. It may be that we are just above the level of the marine limit, and that the lower lying areas nearby, now occupied by swamps contain marine sediments lapping onto the coarser earlier deposited sediments.

76.8 Turn left at Chesterville; continue south over till hill.

77.9 STOP 8. Gravel pit on right along flank of esker (Farmington Falls 7 1/2-minute sheet) A small lake, locally ponded by an ice block in Horseshoe Pond, a till knob north of the pit, and by the Chesterville esker was infilled by pebbly sandy foreset beds, overlain by coarse cobble gravel topset beds.

Return to vehicles, continue south along Chesterville esker.

Kettle ponds occur along either side of the ridge. Several gravel pits in the esker, and in locally ponded material along the flanks of the esker occur. This part of the road traverses the esker crest, and is an excellent example of esker topography for introductory students. Stone (1899) notes the wonderful terminology applied to these features: "In Maine these deposits have received many names. The most common name is 'horseback' .... They are also known as 'whalebacks' and 'hogbacks'. Sometimes one of these ridges is known as the Ridge (as Chesterville Ridge), and they are not infrequently known as 'windrows', 'turnpikes', 'back furrows', 'ridge furrows', 'morriners', and sometimes as 'hills'. Several of these ridges used to be known as 'Indian roads', because Indian trails were made on top of them in the midst of a swampy region. In one place a ridge of this kind was called the 'Indian railroad'. It may be suspected that those who gave it this name had in mind certain archeologists who have thought that the osar ridges were built by Indians."

85.0 Twelve Corners (now apparently only Ten Corners; obviously at town meeting, they voted to throw up the old tote road). Continue south, crossing Route 17.

86.4 Slumped pit on left (Fayette 7 1/2-minute sheet). Gravel pit on left side of road contains large boulders and pebbly sand. Numerous kettle ponds indicate large blocks of stagnant ice occupied the area when the gravel pit sediments were deposited. As long as drainage in the valley to the west of this pit was blocked, streams from the ice margin would have to flow south parallel to the road we are travelling on. However, once the westerly drainage was free, not only would streams most likely follow that route, it would also allow the sea access to the southern portion of the Chesterville valley during subsequent marine incursion.

87.5 Stop sign, turn right toward East Livermore.

88.4 STOP 9. Pit opposite stables (Livermore Falls 7 1/2-minute sheet). Park on left. Walk to excavated area. Laterally continuous planar-bedded silt and silty sand exposed in shallow excavations. These deposits are at approximately 360 feet asl, close to the marine limit, but apparently below effective wave base. These sediments are the best examples of marine deposits seen along the Chesterville valley. Although no marine fossils have been found in these sediments, they occur at just about the level of the nearest marine limit topset/foreset contact 15 miles east in Belgrade (359 feet asl, Thompson and Borns, 1985a). Attig (1975) claims that marine clay occurs on the south end of Jug Hill 0.5 miles northwest of this locality, at an elevation of 394 feet asl. His exposure has not been

found, and no marine clay has been found at this elevation in the Chesterville valley. Smith (1980) and Thompson and Smith (1977) map Presumpscot Formation in this area to elevation of about 360 feet asl; it appears that this exposure we are currently viewing is the best evidence for the marine limit in the area.

END OF TRIP

Return to vehicles; best route back to Lewiston, follow road west to Route 133, turn left (south); follow road to Junction Route 106, turn right (south); follow road to Junction Route 219, turn right (west); - continue on Route 219 to Route 4 at North Turner and take Route 4 south to Lewiston.

TRIP B-8

PETROLOGY AND FIELD RELATIONS OF M2 METAMORPHISM IN WEST-CENTRAL MAINE

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INTRODUCTION

This field trip deals with metamorphism of pelitic rocks, designated M2, in an area which includes Bingham, Anson, Kingfield, and Little Bigelow Mountain 15-minute quadrangles (Fig. 1). Previously M2 has been defined (Guidotti, 1970) in rocks which also experienced effects of M3. Since that time it has been recognized by Holdaway et al. (1982) that M2 is a widespread event being seen as an early metamorphism in the Augusta area (Novak and Holdaway, 1981) and the major metamorphism in Kingfield and Anson quadrangles (Pankiwskyj, 1979) and Little Bigelow Mountain and Bingham quadrangles (Boone, 1973). In the area of this field trip M2 has been less affected by more recent events than in the Phillips and Rangeley quadrangles where it was defined.

Most of the area of study has been previously described by Boone (1973) and Pankiwskyj (1979). The metasediments are mainly Silurian or Devonian and are intruded by gabbros and quartz monzonites of the New Hampshire magma series. The only igneous age available is 400 m.y. for the Lexington batholith (Fig. 1) determined by Gaudette and Boone (1985). Other igneous rocks in the map area are believed to be not significantly different from this age based on their apparent correlation with M2.

The best formations for pelitic specimens are the Silurian Perry Mountain and the Devonian Carrabassett and Seboomook Formations, but most other units have yielded satisfactory specimens on occasion.

This field trip guide represents a progress report on our studies of M2. The sampling detail is greatest in southern Bingham and least in Little Bigelow Mountain quadrangle. Thin sections have been studied from most of the area, but microprobe data are mainly from Bingham quadrangle (Dickerson, 1984).

The Lexington batholith (Fig. 1) has three intrusive phases: a northern coarse equigranular biotite quartz monzonite, a central coarse porphyritic biotite quartz monzonite, and a southern medium equigranular binary quartz monzonite (Boone, 1973; Pankiwskyj, 1979; Koller, 1979). Exposures of the granitic rocks are generally poor. Several phases of basic intrusive activity west and north of the Lexington have been described by Boone (1973).

SUMMARY OF METAMORPHIC EVENTS

The area in question has been affected, at least to some degree, by three metamorphic events as summarized below.

Little Bigelow Mountain
Bingham
1 Chlorite
2 Biotite + chlorite
3 Garnet + biotite + chlorite
4 Staurolite + biotite
5 Andalusite + staurolite + biotite
6 Andalusite + biotite
7 Sillimanite + biotite
A Cordierite + biotite + chlorite
B Andalusite + cordierite + biotite
C Sillimanite + K feldspar + cordierite + biotite
All + muscovite + quartz + ilmenite + graphite
Granitic rocks
Gabbroic rocks
•2 Field trip stop
Northern phase
Central phase
Cord.
Stau.
Stau. Out
Sil.

Figure 1. Preliminary map showing prograde mineral assemblages and isograds for M2 metamorphism in the area of this report. Bar over number indicates highest grade minerals>90% altered. Assemblages 4-7 may include garnet. Intrusive phases of the Lexington batholith are indicated. One inch equals 3 miles.

M1 is widespread chlorite-grade metamorphism which produced S2 schistosity (Guidotti, 1970) prior to 400 m.y. ago. In the study area, S2 is at low angles to bedding, generally strikes northeast, and dips steeply. Subsequent events have been mainly static, producing only local indications of deformation.

M2 is a regional[1]-contact event which produced the generalized sequence chlorite → biotite → staurolite → andalusite-staurolite → andalusite → sillimanite. Most of the sillimanite is clearly contact metamorphic in nature. Cordierite is locally important, taking the place of staurolite. Chloritoid may be seen locally in the biotite zone (Carrabassett Fm.) in northern Anson quadrangle (Fig. 1). Garnet is rarely seen in hand specimen, and in thin section it occurs in small amounts. When seen at low grades it first occurs in the upper biotite zone immediately below the first staurolite. The age of M2 is about 400 m.y. based on the age of the correlative Lexington batholith (Gaudette and Boone, 1985). Much of M2 is directly or indirectly related to the Lexington batholith and other batholiths of the region which produce andalusite in surrounding country rocks (Reddington and Skowhegan).

M3 is a regional-contact event which produced the grade sequence chlorite → biotite → garnet → staurolite → sillimanite. In contrast to M2, andalusite and cordierite are absent from normal pelitic rocks and garnet is common. The zone of sillimanite-staurolite coexistance is narrow relative to the corresponding andalusite-staurolite zone in M2. M3 is mainly developed south of the study area associated with the Hallowell and Livermore Falls groups of plutons, and Phillips Batholith, and the Mooselookmeguntic batholith with ages between 379 and 394 m.y. (Holdaway et al., 1982; Guidotti et al., 1983). Some of the retrogressive effects in the study area may relate to M3.

DEFINITION OF M2

At present there appear to be three necessary elements to any M2 metamorphism: (1) andalusite is widespread in normal pelitic rocks; (2) wherever M3 is also present M2 clearly predates M3 (however in the Augusta area, Novak and Holdaway, 1981, M2 and M3 closely approach each other in time and P-T conditions); (3) retrogressive effects are common except near M2-age plutons. M2 is the only event in this part of Maine which is known to produce andalusite in normal pelitic rocks. As a result andalusite-bearing pelites tend to be assigned to M2.

In addition to the area of Figure 1, M2 is developed in the Norridgewock and Waterville quadrangles, and predates M3 in Augusta, Dixfield, Phillips, Rangeley, and far western Farmington quadrangles. It is interesting to note that M3 rocks of the Livermore Falls group of plutons, exposed primarily in Farmington and Livermore Falls quadrangles, do not show evidence of an earlier M2 event. It appears that this area was first metamorphased at a time late enough that P was too high for development of typical M2 assemblages.

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[1]The use of the term "regional" is intended to imply the absence of a clear relationship to plutons or portions of plutons presently exposed at the surface.

SUB-EVENTS OF M2

Based on the work of Dickerson (1984) near the Lexington batholith M2 may be subdivided into three sub-events which are very closely related in time and partly overlap in P-T conditions:

M2n is contact metamorphism related to the northern phase of the Lexington batholith. The grade sequence is chlorite → biotite → cordierite → andalusite-cordierite → sillimanite-cordierite-K feldspar. Neither staurolite nor almandine garnet are present in these rocks. The rocks are relatively unaltered.

M2 is regional metamorphism which has a very general spatial relation to plutons but may ocur as far as 18 km from exposed igneous rocks. The grade sequence is chlorite → biotite → staurolite → andalusite-staurolite → andalusite, and the rocks are almost always extensively retrograded.

M2s is contact metamorphism around most of the other granitic and gabbroic rocks (including the central an southern phases of the Lexington batholith). M2s overlaps with M2 in time and P-T conditions, but the rocks are invariably less altered than M2 rocks. The grade sequence is the same as that for M2 except that sillimanite without K feldspar tops the sequence. Locally, as in southeastern Little Bigelow Mountain quadrangle, a relatively clear distinction may be made between M2 and M2s (Fig. 1). The two designations are also useful to distinguish between clearly contact-related and regional metamorphic M2, even in places where they were synchronous.

We tentatively assign the age sequence: M2n, M2, M2s. M2n is placed first because it is lowest in P and the metamorphic sequence of the region is clearly one of increasing P with time (Holdaway and Dutrow, in prep.; Dickerson and Holdaway, in prep.). Also, on the west side of the Lexington batholith in the area where northern and central phase activity have both been prominent, there is no sign of cordierite, which would have been produced had the northern phase been intruded last. M2s is considered equivalent and/or slightly later than M2 in age because of the lesser degree of alteration of M2 rocks (e.g. Boone, 1973).

ISOGRAD PATTERN

The isograds which can be traced and which reasonably approximate a univariant P-T line are (1) the first appearance of staurolite or cordierite, (2) the disappearance of staurolite, and (3) the first appearance of sillimanite (Fig. 1). The first appearance of biotite is very composition dependent and requires heavier sampling of a more consistant composition than is available in the area. In M3 rocks the first appearance of Al silicate (as sillimanite) makes a distinct isograd. However in M2, the analogous appearance of andalusite is Fe/Mg composition dependent. In psammitic-pelitic rocks andalusite without staurolite (6)[2] first apears immediately after the staurolite isograd whereas in more pelitic rocks it appears with staurolite (5) at

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[2]Numbers in parentheses refer to assemblages listed in Figure 1.

somewhat higher grades. The uneven distribution of staurolite (4), andalusite-staurolite (5), and andalusite (6) assemblages in Figure 1 illustrates this composition dependence. The disappearance of staurolite is also composition dependent, but an approximate isograd has been shown in Figure 1 based on (1) total absence of staurolite in a variety of rock compositions, and (2) presence of biotite with andalusite (6) in a range of composition (Fe/(Fe+Mg) = 0.46 to 0.70) such that staurolite would be present if grade were lower. As more microprobe work is done the accuracy of this isograd will be improved.

Contact metamorphism is signified by the cordierite isograd for M2n and the staurolite (locally) and sillimanite isograds for M2s. While sillimanite was probably stable along all igneous contacts, a sillimanite isograd is not shown in areas where sillimanite was not specifically sampled. In addition, it is clear from Boone's (1973) and Dickerson's (1984) work that around many gabbroic rocks and around the northern phase of the Lexington batholith the sillimanite zone is less than 100 m wide. One important exception is Little Bigelow Mountain where it appears that the Huston diorite-granite stock (Boone, 1973) passes beneath the mountain to produce regional sillimanite-bearing (7) assemblages (Fig. 1).

Study of Figure 1 shows that regional M2 metamorphism occurs as several "lobes" of medium grade metamorphism, each with one or more plutons within it. One lobe of activity surrounds the Lexington batholith and the neighboring gabbroic bodies. On the south M2 and M2s nearly coincide; on the east they differ by up to 15 km. A second lobe of activity begins along the western edge of Kingfield and Farmington quadrangles. The younger M3 Phillips batholith lies within this lobe. A third lobe of activity may surround the Reddington batholith in the Phillips and Stratton quadrangles (C.V. Guidotti, pers. comm.), and a fourth lobe of activity surrounds the Skowhegan batholith and extends south into the Augusta quadrangle (Holdaway et al., 1982).

Heat sources for these lobes of M2 metamorphism are probably tabular igneous bodies which were in part below or above the present level of exposure (Lux, DeYoreo, Guidotti, and Decker, ms.). Geophysical evidence for a heat source for the complex pattern of staurolite-grade metamorphic rocks east of the southern and central phase of the Lexington batholith does not exist (Mattick, 1965; Kane and Bromery, 1966, 1968). A single gabbro outcrop and abundant gabbro float occur 9 km ENE of Bingham near the north end of the metamorphic high suggesting that the heat source may be thin sheets of gabbroic rocks. The gabboric bodies west and north of the Lexington batholith appear to have little or no affect on gravity for the area (Kane and Bromery, 1968).

Metamorphic reactions and mineral compositions for the Bingham and Little Bigelow Mountain quadrangles are given by Dickerson (1984) and Dickerson and Holdaway (in prep.).

CONDITIONS OF METAMORPHISM

Geothermobarometry in M2 suffers from a shortage of microprobe data at present and from the high degree of alteration of many of the rocks. M3 occurred at conditions where andalusite was no longer stable as a staurolite reaction product. However, the common preservation of M2 andalusite in M3 rocks, and the relatively short time interval between M2 and M3 ($\sim$6 m.y.)

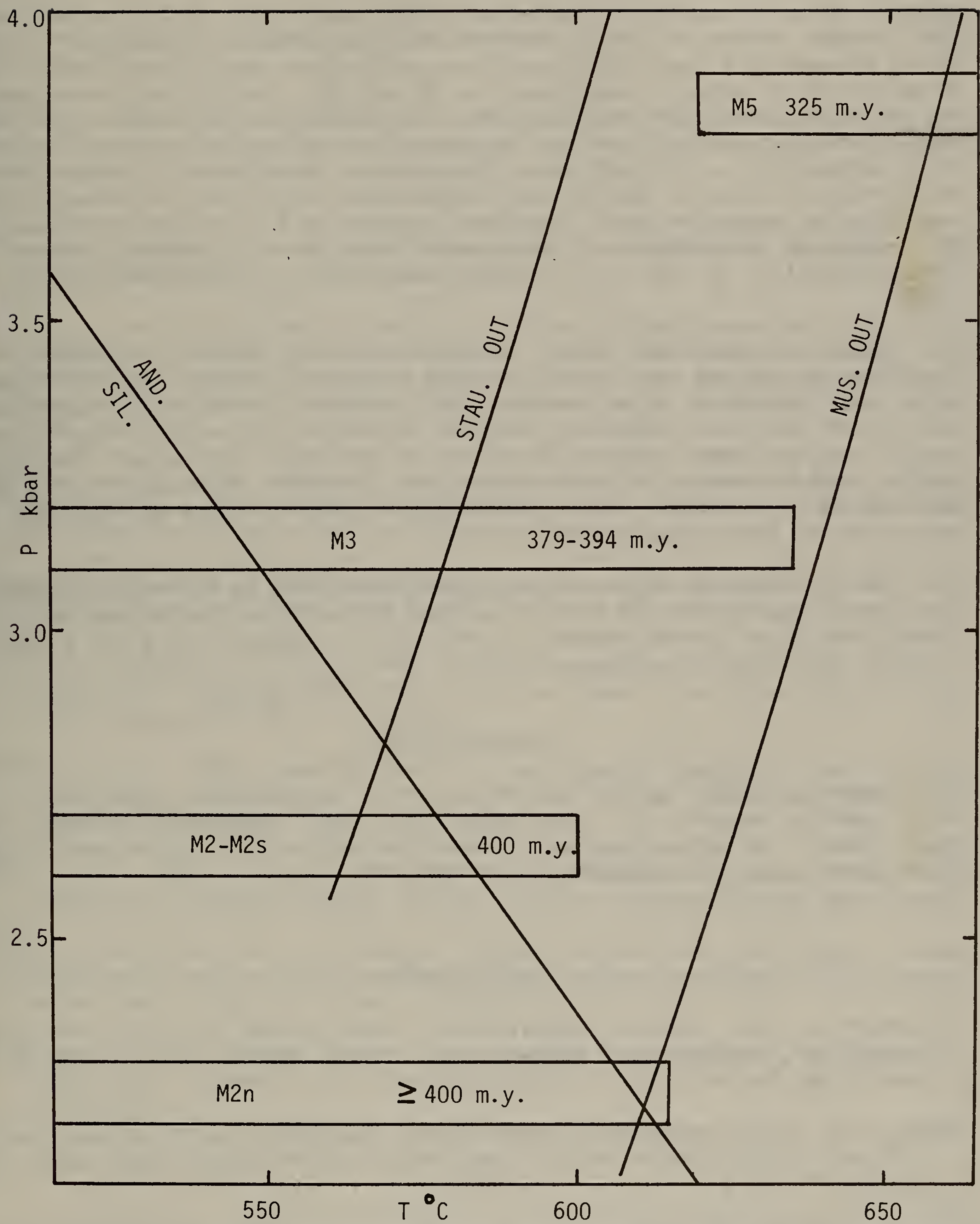

Figure 2. P-T conditions and ages for various metamorphic events in west-central Maine. Sources: Dickerson and Holdaway (in prep.), Holdaway and Dutrow (in prep.), Gaudett and Boone (1985), Aleinikoff (1984).

suggest that M3 conditions passed slightly above the andalusite P-T field. The average garnet-biotite T for the M3 staurolite-out reaction is 580°C. Using reasonable slopes, the staurolite-out reaction may be projected to lower P to provide a range of conditions for M2 and M2s. For M2n muscovite reacted to the first K feldspar and the first sillimanite in the same rocks. Using these data Dickerson (1984) and Dickerson and Holdaway (in prep.) conclude that metamorphic conditions were those shown in Figure 2. There were P differences of several hundred bars between M2n and M2-M2s and between M2-M2s and M3. It should be noted that this increase of P with time continued, and M5 (Hercynian metamorphism) experienced P of about 3.9 kbar north of the Sebago batholith at 325 m.y. (Holdaway and Dutrow, in prep; Guidotti et al., 1986).

This P increase over 75 m.y. throughout a large area of the Central Maine Synclinorium implies that rock was added above the present level of exposure at a faster rate than it was erroded over this time period. Two possibilities to explain this are extensive shallow intrusion and extrusion of igneous rocks, and continued westward thrusting at shallower levels once the present levels were buttressed by metamorphism and intrusion (Holdaway et al., 1982; Guidotti et al., 1983). The P increases may well have been episodic and both processes may have been operative.

The locus of intrusion and related metamorphism in the region appears to have moved toward the SSW with time from M2 to M3 to M5. Thus it is also likely that to a limited extent P increased to the SSW at any given time.

ACKNOWLEDGEMENTS

We are grateful to Charles Guidotti for many helpful suggestions. The high quality mapping of Gary Boone and Kost Pankiwskyj is very much appreciated. We acknowledge with thanks the support of The National Science Foundation, grants EAR-8306389 and EAR-8606489.

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ITINERARY

Assembly point is Arnold's Way Rest Area on U.S. Route 201, east side of road, 2.3 miles north of downtown Solon (1.75 hour drive from Lewiston). Note change from original announcement. Assembly time is 9:15 A.M. Topographic maps: Bingham, Anson, Kingfield, Little Bigelow Mountain, and Farmington quadrangles. Only the key mineral(s) of assemblages are given in the guide. Number in parentheses designates complete assemblages as given in legend to Figure 1.

Mileage

0.0	Proceed north from Rest Area on U.S. Route 201 and enter Bingham quadrangle at 1.2 miles.
1.5	Roadcuts of purple and light green layered Madrid Fm.
3.3	Tooth and Claw Jewelry Shop on left.
3.5	Turn right (east) on Mahoney Hill Rd.
4.0	Discontinuous roadcuts of Madrid and Carrabassett Fms. represent decreasing grade from andalusite (6) to biotite (2).
5.8	STOP 1: Johnson Brook, south of road. Walk 100 m downstream to outcrops of possible psammitic Madrid Fm. and pelitic Carrabassett Fm. This represents the southwestern limit of a closed M2 high which extends 3.5 miles NE. Pelitic rocks contain staurolite (4) whereas chlorite-bearing psammitic rocks contain neither staurolite nor andalusite. Continue east on Mahoney Hill Rd.
6.5	STOP 2: Just beyond top of the hill on Johnson Mountain. Typical staurolite schist (4) of Carrabassett Fm. Much of the staurolite is altered to fine sericite and chlorite. One mile NNE of here is chiastolite-staurolite schist (5) which extends for a mile along the top of Johnson Mountain, making the core of the eastern high. Turn cars around and return to U.S. Route 201.
9.5	Turn right (north) on U.S. Route 201.
10.3	Bingham city limit.
11.4	Turn left (west) on ME Route 16 across the Kennebec River.
11.6	After bridge, turn right (north). Massive rusty roadcuts are Smalls Falls Fm., mainly sulfidic-graphitic and psammitic rocks.
13.3	Wyman Dam. Around bend to north metapelites of the Perry Mountain Fm. are retrograded andalusite-staurolite schist (5).
13.7	Perry Mountain Fm. contains retrograded andalusite-staurolite schist (5). Rusty Smalls Falls roadcuts begin.
15.0	Trail to Houston Brook Falls on right, in psammitic rocks of Madrid Fm.
15.8	Take left (northwest) fork toward Rowe Pond.
17.5	Longfellow School.
17.8	STOP 3: Low outcrops on left. Slightly altered staurolite schist (4) of Carrabassett Fm. (Andalusite may occur locally). The M2 staurolite isograd crosses the road (in Madrid rocks) several hundred meters north of here and trends WNW toward the junction between the northern and central phases of the Lexington batholith.

18.05 Small outcrop of staurolite schist (4) on left.

18.15 Small outcrop of biotite psammite in Madrid Fm. on left.

18.4 Pavement ends. Continue north.

19.8 Low outcrops on left are Carrabassett Fm. cordierite hornfels (A) of M2n.

20.1 STOP 4: Stop at road to south end of Clear Pond. Cordierite hornfels (A) of M2n in Carrabassett Fm. Glaciated outcrop on right before road and better outcrop and float for collecting on left 0.1 mile after road. Most or all of these rocks contain only cordierite (A), but rocks on the hill 0.5 mile WSW also contain 2 x 15 mm sharply defind andalusite with the cordierite (B). These assemblages, without staurolite, occur on the east and north sides of the northern phase, Lexington batholith. On the west side, later gabbroic rocks and central phase intrusives have destroyed the cordierite. Continue north.

20.4 STOP 5: Rubbly outcrop of northern phase coarse biotite quartz monzonite. This rock differs distinctly from the porphyritic central phase. Continue north.

20.6 Turn cars around. Watch for cars coming over hill. Return to U.S. Route 201 in Bingham.

25.4 Turn right (south).

29.6 Turn left (east) on ME Route 16 across Kennebec River.

29.8 Turn right (south) on U.S. Route 201.

31.8 Mahoney Hill Rd.

34.0 Enter Anson quadrangle.

35.2 LUNCH STOP: Arnold's Way Rest Area.

36.6 Solon city limit.

36.9 Turn right (west) on Falls Road.

37.2 Continue straight on dirt road.

37.3 STOP 6: Park on dirt road to left. Arnold's Landing. CAUTION: dam gates may be opened without warning. Staurolite (4), commonly rimmed by alteration products, dominates in pelitic Carrabassett layers, whereas coarse andalusite pseudomorphs (6) occur without staurolite in the more psammitic (Fall Brook Fm.?) layers. Return to U.S. Route 201.

37.6 Turn right (south) on U.S. Route 201.

38.2 Psammitic Fall Brook Fm. in stream to left.

38.3 Turn right (east).

38.8 Take right fork on Athens Rd.

39.0 STOP 7: Park on top of Rise. Roadcuts on left are slates of the Carrabassett Fm. which contain chlorite and no biotite (1). Note that the distance separating stops 6 and 7 is 1.2 miles. Turn cars around and return to the center of Solon.

39.7 Cross U.S. Route 201 and continue west on U.S. Route 201A across Kennebec River.

40.9 Continue left (south) on U.S. Route 201A.

41.4 Turn right (west).

42.2 New roadcuts in Fall Brook Fm. contain bioite (2) in pelitic layers.

42.5 Power line.

43.4 Roadcuts of andalusite-staurolite schist (5) in Carrabassett Fm.

43.9 Cross Dunbar Hill Rd. Carrabassett Fm. andalusite-staurolite schist (5) 0.1 mile south, Fall Brook Fm. and Perry Mountain Fm. andalusite schist (6) north and northwest of here.

45.6 Turn left (south)

45.7 Turn right (west) on Wentworth Rd.

46.9 Outcrops in this area are psammitic Fall Brook Fm. 0.9 mile northwest of here is andalusite schist (6) in Perry Mountain Fm., 0.4 mile from the southern phase of the Lexington. M2 and M2s merge here (Fig. 1).

48.3 Turn right (north) on ME Route 16.

49.5 Roadcuts of Fall Brook and Smalls Falls Fms. Madrid Fm. hornfels at East New Portland (2 miles west of here) contains diopside.

49.8 Road to East New Portland on left. Continue north on ME Route 16.

50.7 Enter Lexington batholith, southern phase.

50.9 Enter Kingfield quadrangle.

51.3 STOP 8: Roadcuts of southern phase, Lexington batholith on both sides. The rock is a medium-grained binary quartz monzonite with a high inclusion content at this locality. According to Koller (1979) there is no sharp boundary between the southern and central phases. Although sillimanite occurs in country rock close to the igneous rock contacts, appropriate compositions are rare around the southern phase (Fig. 1). Continue northwest on ME Route 16.

52.6 Sharp left (west) in North New Portland. Stay on ME Route 16.

52.7 Granitic outcrops 100 m left of bridge on east side Gilman Stream.

57.3 Leave Lexington batholith.

57.5 Roadcuts of psammitic Madrid Fm.

58.5 On left, glacial boulder of coarse, porphyritic biotite quartz monzonite of the central phase, Lexington batholith. Microcline phenocrysts are up to 10 cm long and commonly define a flow pattern.

60.2 After crossing Carrabassett River in Kingfield turn right (north), staying on ME Route 16. Psammitic Madrid Fm. below bridge.

62.0 River outcrops of Carrabassett Fm. andalusite schist (6) which occur discontinuously nearly to stop 9.

63.3 Enter Little Bigelow Mountain quadrangle.

66.5 Roadcuts in Seboomook Fm.

66.9 STOP 9: Roadcuts show large, sharply defined andalusite pseudomorphs in Seboomook andalusite schist (6). The higher aspect ratio and better crystal face development are more characteristic of andalusites close to gabbro or granite contacts. These characteristics are apparently not a function of lithologic unit; Carabassett schists 0.5 mile up Hammon Field Brook (in this same area) have the same texture. Andalusites closer to igneous contacts are commonly smaller. West of the highway are staurolite schists (4) in these M2 rocks. To the east andalusites eventually become less altered, (M2s) and, within 0.2 mile of the granite, relatively fresh sillimanite hornfels (7) occurs (Fig. 1). Continue north to Rest Area.

67.0 Carrabassett Valley Rest Area. Turn cars around and head south.

68.4 STOP 10: Roadcuts in Carrabassett Fm. show smaller, more sievy, and less well defined andalusite pseudomorphs (6). This character is comparable to that of other regional M2 andalusites. Continue south.

74.0 Turn right (west) on ME Route 142 in downtown Kingfield. Route 142 lies mainly on glacial outwash.

77.6 Turn left (south) on ME Route 145. This point is the junction between two lobes of M2. Northwest of here are biotite-grade (2) rocks. To the southeast, Freeman Ridge belongs to the Lexington lobe; to the southwest Foster Hill belongs to the western lobe which becomes important in Phillips and Dixfield quadrangles. The route south winds between garnet schists (3), staurolite schists (4), and andalusite-staurolite schists (5) of this western lobe. (Fig. 1).

81.1 Road to Salem. Hill 1.2 mile northwest contains andalusite-staurolite (5), and hill 1 mile southeast contains staurolite (4), both in Seboomook Fm. Continue south.

84.7 Stream outcrops on left in Seboomook Fm. contain garnet and biotite (3) which have been totally replaced by chlorite.

87.3 Junction with ME Route 149 in Strong. Continue on Route 145.

87.7 Junction with ME Route 4. Proceed south on 4.

89.6 Roadcuts on right are Carrabassett Fm. biotite schist with occasional retrograded garnets (3).

90.0 STOP 11: Roadcuts on right. Carrabassett Fm. retrograded staurolite-garnet schist (4) in more pelitic layers. Rocks to the east are mainly chlorite grade (1). Along the western side of Kingfield and Farmington quadrangles garnets are more common in M2 and a garnet zone is mapable at least in the Farmington quadrangle (Dutrow, 1985). This and a wider M2 staurolite zone in Farmington and Dixfield quadrangles suggest that here M2 may have occurred at slightly higher P than near the Lexington batholith.

91.5 Enter Farmington quadrangle. Between here and Farmington most road-cuts are quartzitic and sulfidic-graphitic compositions.

96.2 Intersection with ME Route 27. Continue south on ME Route 4 through Farmington.

99.3 Intersection with U.S. Route 2. Continue south on Routes 2 and 4.

99.9 Roadcuts of garnet schist of the Sangerville Fm. Rocks in the area south of Farmington have suffered M2 and M3 garnet-zone conditions. Dutrow (1985) has observed two stages of garnet growth.

101.3 Roadcut on left of M3 staurolite-garnet-biotite schist, Sangerville Fm.

102.3 Intersection with ME Route 133. Continue south on Routes 2 and 4.

104.3 STOP 12: Long roadcut on left. These Sangerville Fm. staurolite-zone M3 rocks are provided for comparison with M2. The mineral assemblage of pelitic rocks is staurolite-garnet-biotite-muscovite-quartz with minor chlorite alteration. The M3 isograds trend WSW while the M2 isograds trend approximately south. Retrograded M2 staurolites are seen near Temple and Varnum Pond in west-central Farmington quad-rangle, and the last remnants of M2 disappear north of Wilton. The M3 rocks of Farmington quadrangle have their heat source in the Livermore Falls group of plutons which occupy much of the southern 40% of the quadrangle. The Farmington M3 rocks are unusual for M3 in that most of them apparently did not experience an earlier M2 event (except along the western side). M3 rocks elsewhere have occasional coarse andalusite remnants from earlier M2. Along Route 4, the first M3 sillimanite appears, without andalusite, in large roadcuts immediately south of Wilton. Continue south on ME Route 4 to Lewiston-Auburn, about 40 miles.

TRIP B-9

QUATERNARY GEOLOGY OF THE DAMARISCOTTA ESTUARY

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INTRODUCTION

Maine is popularly known for its "rock-bound coast." This is an apt description for the exposed marine littoral, but there are a multitude of other environments in more protected locations. The coast of Maine is more accurately characterized by peninsulas, islands and embayments. Pleistocene till, outwash, and glaciomarine mud form bluffs which are a source of sediments to the Holocene systems. Within the headward and marginal portions of embayments are broad subtidal and intertidal flats, fringing salt marshes, and occasional broad marshes. The relative abundance of the rocky ledge, mudflat, marsh, and bluff environments varies along the coast, depending on bedrock structure and Quaternary history. The coastal and Quaternary setting of Maine's coast has recently been summarized by Kelley and Kelley (1986).

Coastal Environments

The Maine coast has been divided into four coastal geomorphic compartments (Jackson, 1837; Kelley, in press) (Figure 1). The divisions of this classification reflect underlying controls by bedrock structure and Quaternary sediment supply. The southwest coast (SW), from Kittery to Cape Elizabeth consists of arcuate pocket beaches and barrier spits as well as rocky headlands. This zone has extensive broad marshes behind the barriers and in drowned river valleys. Abundant sandy outwash has allowed the development of these Holocene barrier systems.

The west-central compartment (WC) from Cape Elizabeth to Rockland is a drowned coast of long linear embayments and peninsulas. The form of the coast is controlled by underlying bedrock structure and lithology subjected to differential erosion. This compartment can be further subdivided by the orientation of the estuary with respect to bedrock strike: strike parallel embayments, such as the Damariscotta River, and strike-normal embayments, such as Casco Bay (Belknap et al., in press 1986b). This distinction is related in part to orientation of glacial flow with respect to bedrock strike, and resulting glacial scour and glacial and glaciomarine deposits. During the latest Pleistocene deglaciation and marine submergence, this area was a deep embayment, and was covered with thick glaciomarine mud, the Presumpscot Formation of Bloom (1963).

The east-central (EC) compartment is controlled by large granitic intrusions which form islands. The resultant geomorphic form is islands, bays

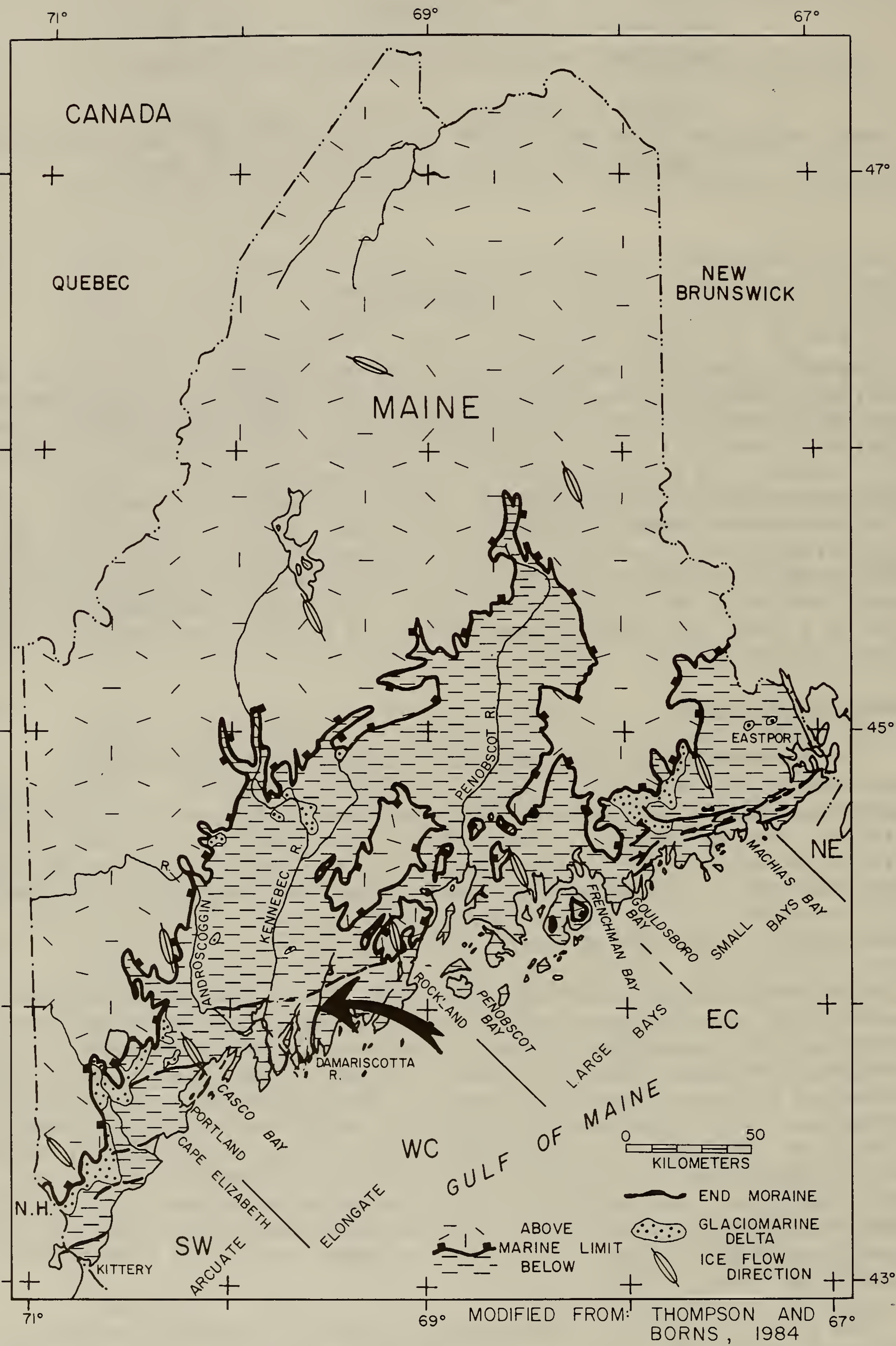


Figure 1. Location Map. Maine coast with coastal compartments, and the extent of the marine submergence. After Belknap et al., 1986b.

and headlands. Shipp et al. (1985) have further subdivided this compartment into the large bays of Penobscot Bay to Frenchman Bay and small bays from Gouldsboro Bay to Little Machias Bay.

The northeast (NE) coastal compartment is a straight cliffed coast of metavolcanic and metasedimentary rocks, with a few pocket beaches. Its shape is controlled by Paleozoic bedrock faults and wave activity.

Quaternary Sea-level Changes

The coast of Maine has been affected strongly by relative sea-level changes over the past 13,000 years. During deglaciation, the sea was in contact with retreating ice, and reached 70-80 m above present in the Damariscotta region about 12,500 years B.P. (Figure 2). At this time there was a short high stand as ice grounded and glaciomarine deltas were deposited (see: Thompson and Borns, 1985). From 13,000 to about 11,000 yrs. B.P. the glaciomarine Presumpscot Formation was deposited over the region. As glacial ice retreated, the rate of isostatic rebound outpaced eustatic sea-level rise, and so relative sea level fell rapidly to a lowstand at 65 meters below present about 9500 yrs. B.P. During lowstand Pleistocene sediments were subjected to subaerial and coastal erosion and significant shoreline features were produced. Finally, as the rate of isostatic rebound decreased, relative sea level rose with eustatic rise. The curve is best constrained by ^{14}C dates on salt marsh peats for the period after 6000 years B.P. These sea-level changes are described in more detail by Belknap et al. (in press 1986 a,b,c,d), and by Schnitker (1974a). The changing levels of land and sea thus resulted in a marine submergence in immediate contact with the ice, followed by a rapid regression, a significant lowstand, ultimately followed by the rapid and then slowing transgression which continues today.

The deglacial and Holocene sedimentary processes have been strongly affected by these sea-level changes. Analysis of over 3000 km of high resolution seismic profiles, over 110 vibracores, and over 500 marine sediment grab samples has allowed us to model the geologic history and stratigraphy of these Quaternary marine units. A typical stratigraphic section for the coast is based on glacially-sculpted bedrock, capped by glacial and deglacial sediments. The base of the sedimentary section is often diamicton, either a thin basal till or mounded into moraines of various sizes. Overlying the till, and interfingering with the moraines in some areas is the glaciomarine mud of the Presumpscot Fm. Also interfingering with the Presumpscot Fm. is stratified sand and gravel outwash, in some places accumulating as glaciomarine deltas. The Pleistocene units are capped by an erosional unconformity, notable for shorelines, fluvial erosion, and soil development. The top of the Presumpscot Fm., in particular, is notable for an oxidized, desiccated zone of subaerial weathering, for gullies, and for slumps. Overlying the unconformity are Holocene estuarine and marine deposits, most commonly consisting of tidal flat mud overlain by more marine mud as transgression continued. In some places salt marsh peat has been preserved. At the mouth of major rivers, such as the Kennebec, sand and gravel was reworked from upstream glaciofluvial sources (Borns and Hagar, 1965) and has been deposited as deltaic deposits, now submerged below sea level near the lowstand (Oldale et al., 1983; Belknap, 1985).

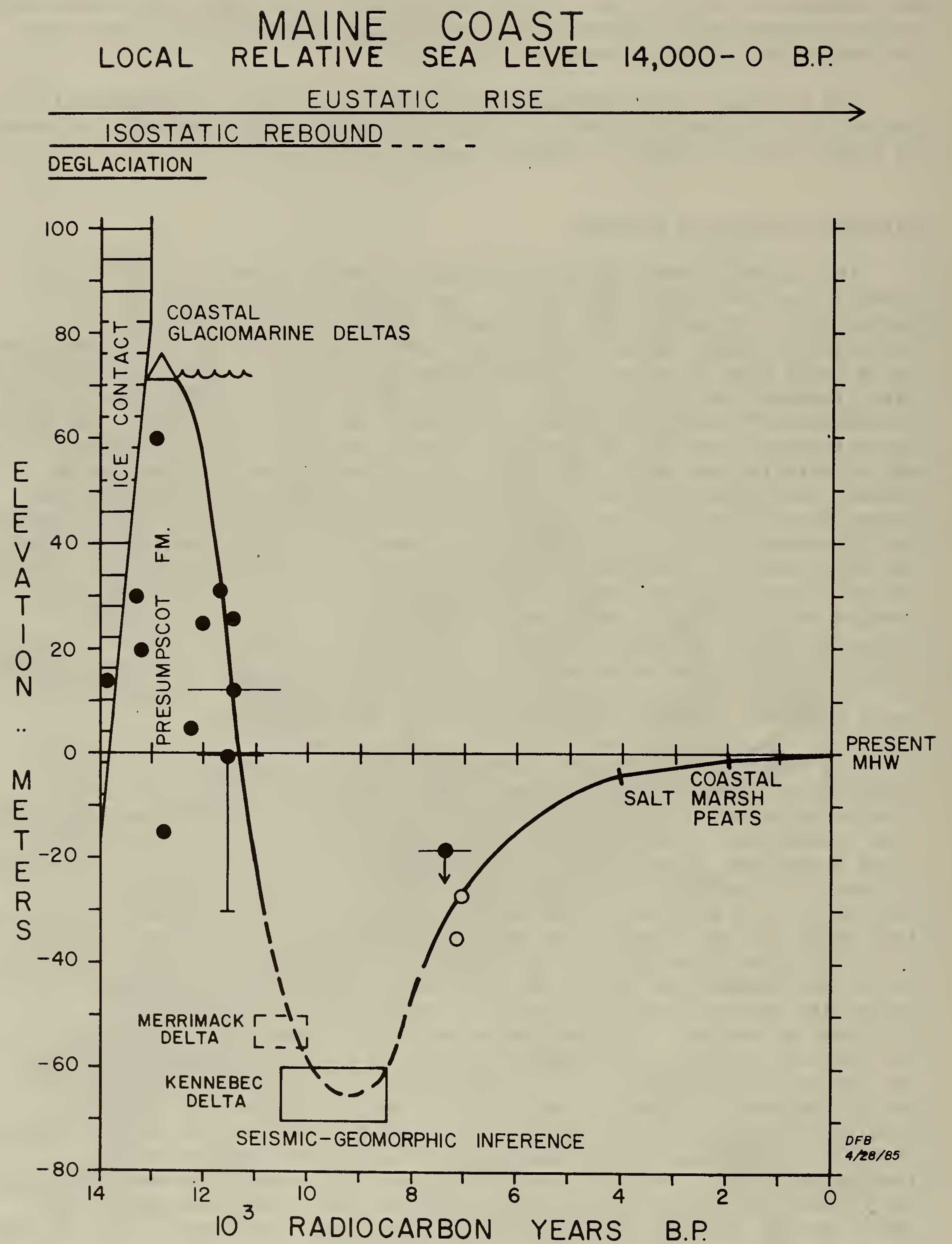


Figure 2. Late Quaternary local relative sea-level curve, based on radiocarbon dated shells and peats and present elevations. After: Belknap et al., 1986b.

Holocene Coastal Evolution

During the late Holocene the Maine coast has evolved through a combination of rising sea level, coastal erosion, and sediment accumulation in protected locations. A model of this evolution (Figure 3) has been presented by Kelley (1986 in press) and Belknap et al. (1986b in press). The estuarine model is divided into three zones.

In the inner zone (zone I) marsh and tidal flats are the dominant sedimentary environments, with vegetated, stable bluffs on the landward margins of marshes, and unstable bluffs often found behind flats where marshes are absent. Turbidity is highest in this section of the estuary. This zone is considered to have rapid rates of sediment accumulation (1 cm/year or greater, as determined from ^{210}Pb, ^{137}Cs, and pollen data: Schnitker, 1972; Hay, in preparation). This sediment may come from the rivers, from the Gulf of Maine (Schnitker, 1974b), or most likely from internal recycling of Pleistocene and Holocene sediments (Belknap et al., 1986a,b in press; Kelley, 1986 in press). In the central estuarine zone (zone II) marshes are small, discontinuous and only fringe the embayment. Intertidal and subtidal mudflats are more common than marshes, and often front eroding bluffs. Sediment accumulation is episodic, primarily from slumping of bluffs, and highly variable. Anderson et al. (1981a,b) measured highly variable rates of sedimentation and erosion which were seasonally linked (erosion from October to March) in this zone in the Damariscotta River estuary at the protected Lowes Cove, but computed a net accumulation rate of 0.9 cm/yr.

The marine end of the estuary (zone III) is stripped nearly bare of sediments, both in the intertidal zone and on the rocky subaerial peninsulas. The only sedimentary environments are high energy pocket beaches with gravelly flats in protected coves. Below wave base in the estuary, however, portions of the stratigraphic column are preserved. Mud flats are winnowed at the interface of zones II and III, producing sand and gravel lags while the mud is moved landward and seaward by tidal currents.

The predictions based on this model are that as sea level rises, these three zones are displaced landward, in accordance with Walther's Law. The degree of preservation of sediment from any specific environment depends on estuarine geometry, which controls the local importance of each environmental lithosome, and the rate of sea-level rise, as discussed by Belknap and Kraft (1981, 1985). The sediments are recycled from seaward (zone III) to landward (zone I), storing sediment temporarily in the flats and marshes. New sediment is introduced primarily from bluff erosion, with a low amount of primary fluvial input, and an unknown component of marine sediments brought in by estuarine circulation. This model applies to the present slowly rising sea level. It is likely that at a lower stage of sea level, with more rapid rise, sediments may have been efficiently carried out of the estuaries and dumped into the deeper marine basins.

The Damariscotta River Estuary

The Damariscotta River (Figure 4) is a typical example of the WC coastal compartment with a long, narrow, drowned valley flanked by rocky peninsulas.

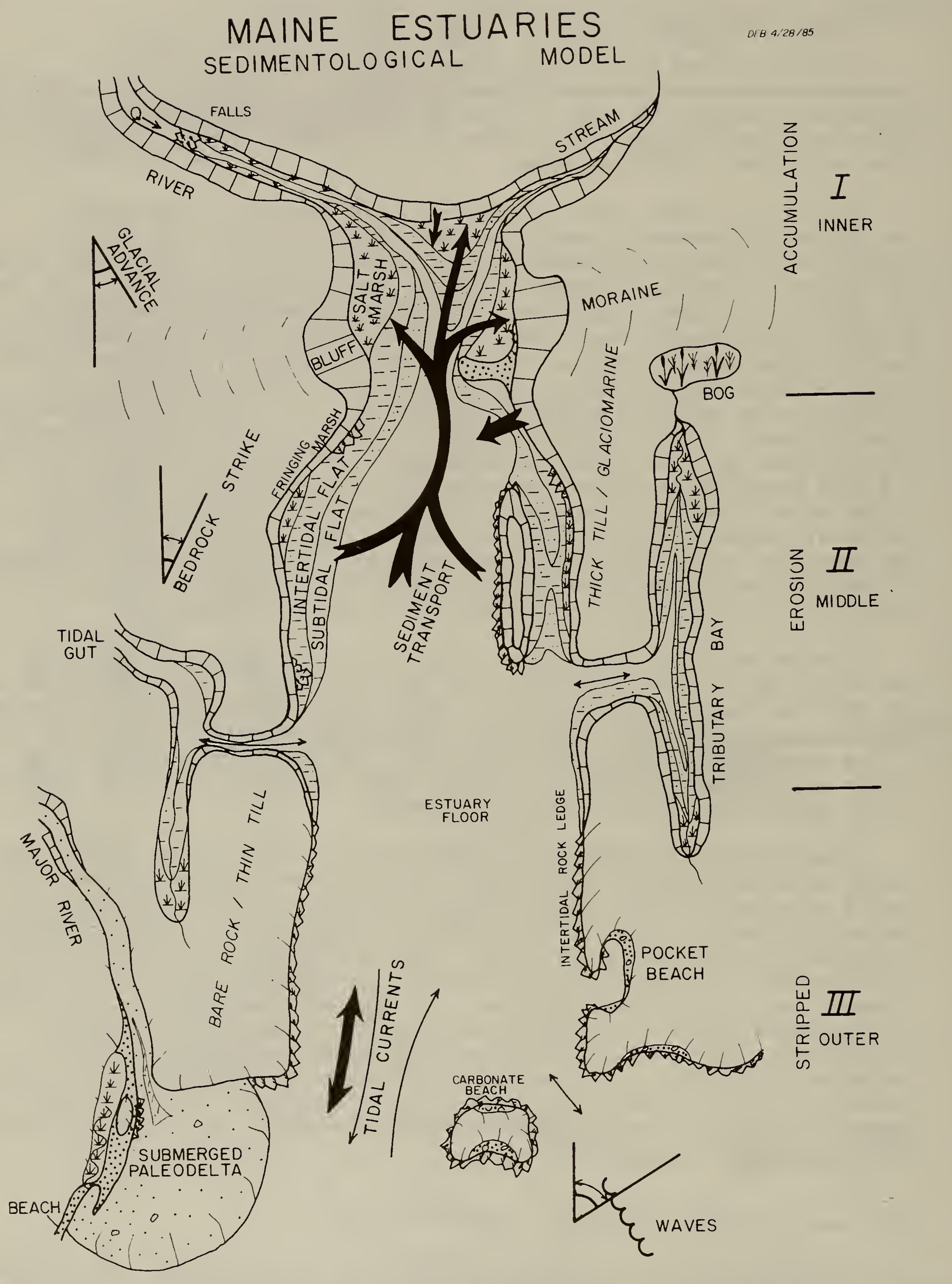


Figure 3. A model of Holocene estuarine evolution for Maine, after Belknap et al., 1986b and Kelley, 1986. Postulated sediment transport pathways shown by the broad arrows.

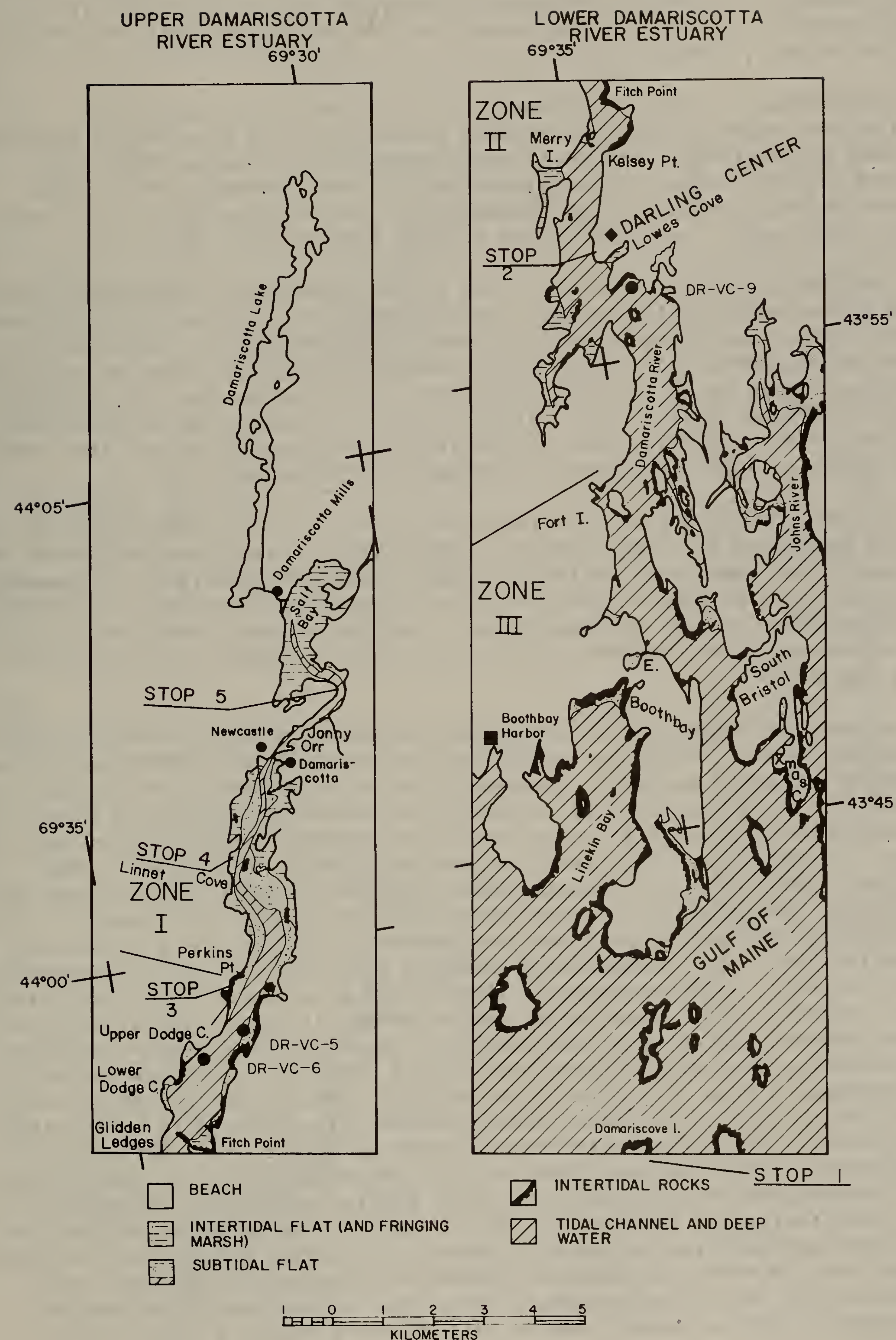


Figure 4. Location map of the Damariscotta River estuary, with place names used in the text, and stop locations.

It is 25 km long and averages about 1 km wide with some constrictions and wider reaches. The axis is sinuous to straight, and nearly parallel to bedrock strike. It is the type example of the strike-parallel embayment (Belknap et al., 1986a).

The bedrock is highly deformed schists of medium to high-grade amphibolite facies. Their sedimentary precursors were interbedded sandstones and impure limestones. The rocks are identified as Devonian-Ordovician Bucksport Formation on the new state geologic map (Osberg et al., 1985). Because of its orientation, the Damariscotta estuary lies almost entirely within this single rock type in the coastal lithotectonic block.

Bathymetrically, the estuary is a series of semi-enclosed basins, with sills and reversing falls at the constrictions. At the mouth, it is primarily an open marine embayment up to 37 m deep, while above Fitch Point it is estuarine and dominated by sub- and intertidal flats with some marshes. The mean tidal range is 2.7 m, 3.3 m at spring. Freshwater input is from the dammed Damariscotta Lake, at a natural falls, at a mean rate of 3 m^3/sec. McAlice (1977, 1979) has shown from extensive oceanographic data that the Damariscotta is usually a type B, partially mixed estuary.

The estuary is not heavily impacted by man, with open pastures and heavily wooded margins most common. Pollution of tidal flats and water is limited to the immediate proximity of Damariscotta-Newcastle. In the past, the river has been used for shipbuilding in numerous locations, an industry which continues in East Boothbay today. Another formerly important industry was brickmaking, the remnants of which can still be seen at many locations along the shore.

Seismic stratigraphy for the estuary is based on over 175 km of high-resolution seismic reflection profiling, using Raytheon RTT1000A 3.5 kHz and ORE-Geopulse boomer systems. Interpretation of this remote sensing data is constrained by outcrops and by 12 submarine vibracores. Figure 5 is a compilation of typical 3.5 kHz records from the upper, middle, and lower estuary. Bedrock is recognized by a distinct strong return, while the localized examples of till give a deeper penetration with chaotic internal reflectors. Minor moraines are evident in several locations, usually correlatable with outcropping moraines onshore. Pleistocene glaciomarine sediments give a sharp return with internal reflectors or more transparent underlying units. In the Geopulse record, the glaciomarine units are seen to have a rhythmic bedded appearance, which changes from a draped geometry to a ponded geometry moving up in the section. This is interpreted as a change from deepwater proximal facies of very rapid over- and interflow deposition from suspension to a more distal, but shoaling sequence dominated by turbid underflows. Similar interpretations have been made in Canadian estuaries by Piper et al. (1983). The glaciomarine deposits are cut by a distinct unconformity, and often show truncation of reflectors (e.g. Line D-16, Fig. 5). The thalweg of the estuary is often barren of Holocene sediments, indicating continuing erosion by tidal currents. The margins, on the other hand, have accumulations of several meters of Holocene mud. Some deep basins contain over 10 meters of Holocene estuarine deposits. In general, the pattern of the seismic stratigraphy is similar to the modern sedimentary environments, with an outer zone of thin Holocene units, a middle zone

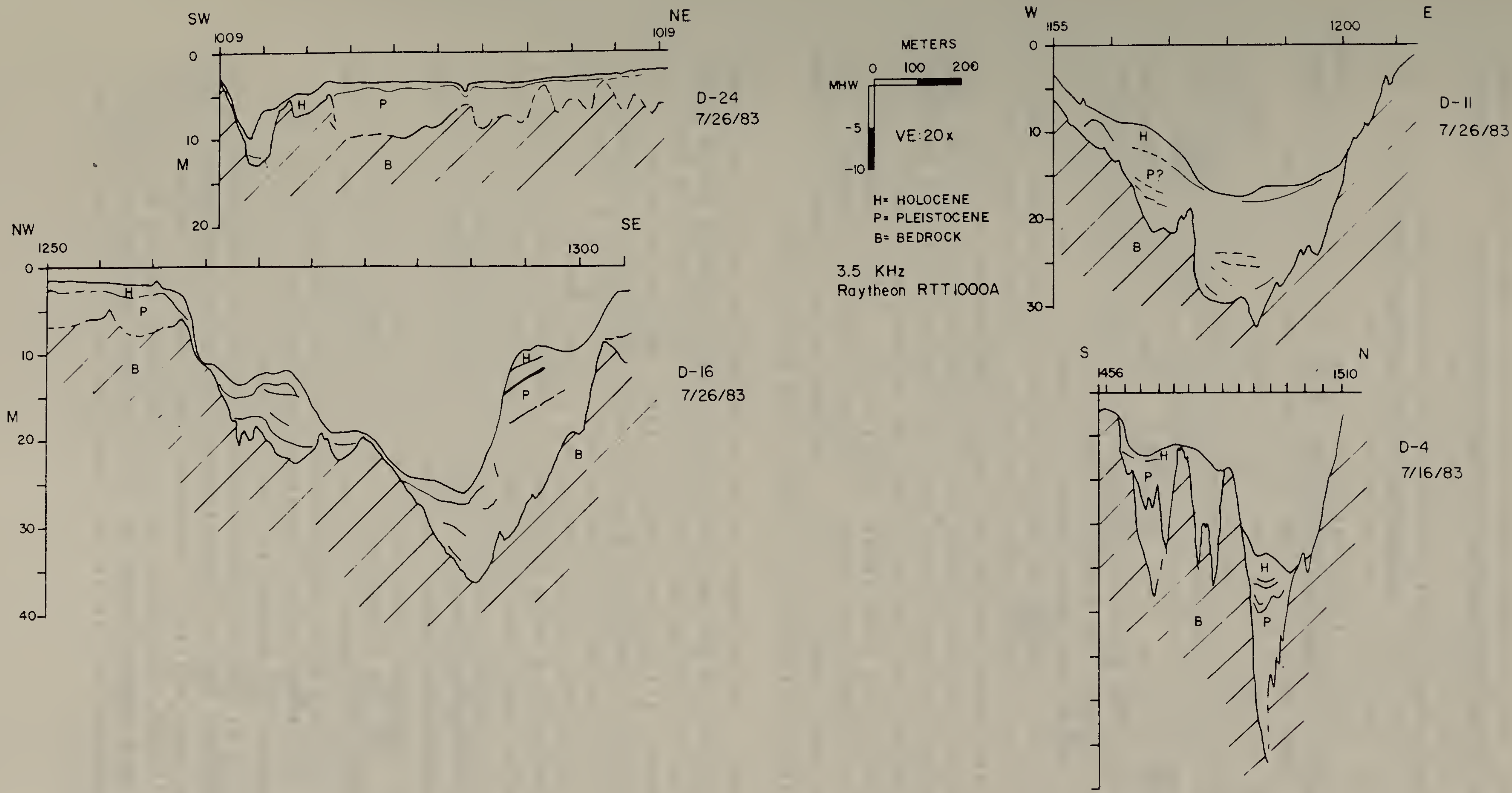


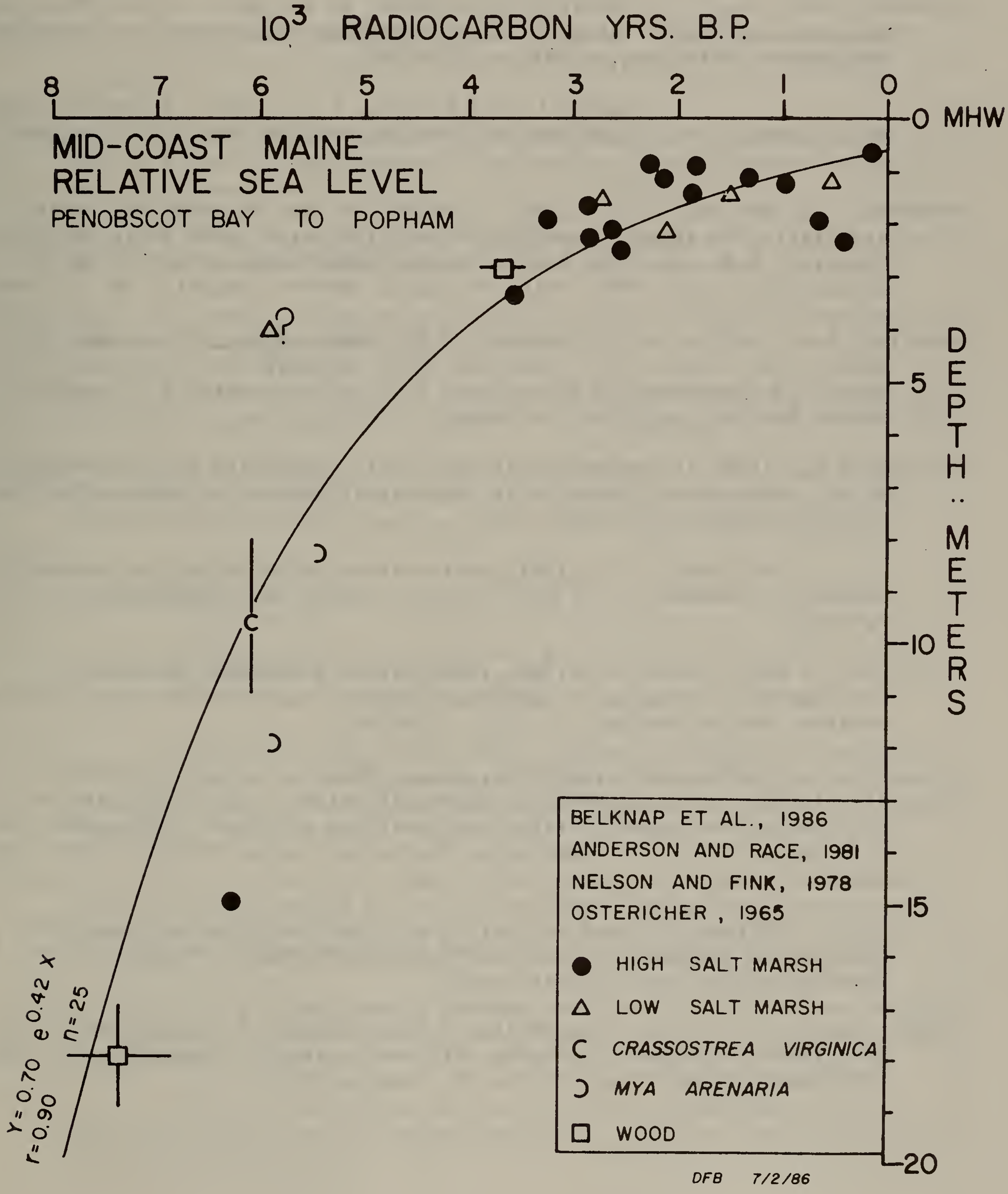
Figure 5. Line drawings of typical 3.5 kHz high-resolution seismic reflection profiles within the Damariscotta River. The lines are approximately normal to the river axis, and are arranged from top to bottom from inner to outer estuary (Figure 4). Line D-24 is located approximately 1 km south of Linnet Cove. Line D-16 is located at Dodge Lower Cove. Line D-11 is located at Kelsey Point. Line D-4 is located approximately 0.5 km south of Lowes Cove.

undergoing redistribution, and an inner zone of sediment accumulation. Thus, the model applies to the three dimensional package of sedimentary lithosomes.

The Holocene local sea-level curve for mid-coast Maine (Figure 6) is based on peats and shells from cores in marshes and the deeper embayments. The best-fit curve through the data is an exponential curve, with a correlation coefficient of 0.90. The curve shows a rapid rise of 5.45 m/1000 yrs. about 7000 years B.P., when sea level was 14 meters below present, slowed to about 1.2 m/1000 yrs 3000 years B.P. when sea level was 2.5 m below present, and finally has slowed to less than 0.5 m/1000 years over the past 2000 years. A further constraint on late Holocene sea-level rise is provided by the Damariscotta shell heaps immediately south of the Route 1 bridge. The base of these Crassostrea virginica oyster middens has been dated at about 2300 yrs. B.P. (Sanger and Belknap, 1986 in press). They provide an important clue to sea-level rise, as they occupy a tidal river which is above a bedrock sill now flooded by about 1 meter of water at mean high tide. This level is consistent with the salt marsh peat data shown in Figure 6. The first appearance of the oysters, and subsequent exploitation by Amerindians thus approximately dates the first flooding over that sill. Other middens are common farther south in the Damariscotta, but they are composed almost exclusively of Mya arenaria, the soft-shell clam. Anderson et al. (1984) have shown evidence for a recent more rapid rate of rise, particularly in eastern Maine surrounding the Eastport region, which may be related to crustal warping. Data from the Portland tide gauge (Hicks et al., 1983) suggest an ongoing rise of 2.3 mm/yr (m/1000 yr): faster than any local long-term Holocene rate later than about 5000 yrs. B.P. Belknap et al. (1986c in press) discuss Maine's local relative sea-level changes in greater detail.

In summary, we present the following model of estuarine evolution, based on our data and extrapolation to the earlier part of the Holocene where data is sparse. The geological evolution of the Damariscotta estuary has been dominated by late Quaternary glaciation, deglaciation, and relative sea-level changes. As the ice retreated, bedrock, glacial and glaciomarine sediments were exposed to marine processes of erosion as sea level fell to a lowstand 65 m below the present. A deep valley was cut in the present Damariscotta, an the outer parts of the peninsulas were stripped of sediment. As sea level rose and drowned the valley once more, Holocene sedimentary environments took on an appearance similar to today's, with flats and marshes dominant in the upper reaches of the protected embayment. Early in the transgression, when sea level was rising rapidly, sediments reworked from bluffs and uplands were efficiently carried into the deeper marine, with some proportion carried headward in the estuary. As the transgression continued, a series of sills and basins were overstepped and opened to estuarine conditions. Some of these may have been lakes, like the modern Damariscotta Lake and Biscay Pond - Pemaquid Pond systems. The most recent overstepping occurred at Johnny Orr, the bedrock sill in Damariscotta-Newcastle, with a resulting explosive population growth of oysters, and subsequent exploitation by Amerindians. As the transgression flooded the valley, Pleistocene bluffs and earlier Holocene sediments were eroded by wave activity, slumping, and tidal currents. As the rate of sea-level rise slowed, sediments were increasingly stored within the estuary, in zone I. In the present setting, we find most sediments to be locally derived and deposited, and little evidence for major influx from or efflux to the deeper marine.

Figure 6. Holocene local relative sea-level curve for central coastal Maine, from Penobscot Bay to Popham. Based on data from published reports and Belknap et al. 1986c in press. The best-fit curve is an exponential function on all the data points except the queried low-marsh peat from Popham.



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ITINERARY

Assembly point for this trip will be in front of the administration building on the grounds of the I.C. Darling Center of the University of Maine, in Walpole, Maine at 0900. A sign along Route 129 approximately five miles (8 km) south of Damariscotta marks the turn to Clark's Cove Road on which the laboratory is situated. Topographic maps that cover the field trip are the Damariscotta, Bristol, and Pemaquid Point 7.5′ quadrangles. Nautical chart #13293 from the National Ocean Survey covers the entire area at a scale of 1:40,000.

From the Darling Center, participants will board the shuttle vans and be driven to Christmas Cove, located five miles (8 km) south of the Darling Center. At the Coveside Marina dock everyone will board the research vessels R/V *Lee* and R/V *Miss Bess* of the University of Maine fleet for the trip to the mouth of the Damariscotta River. Because of the size of the vessels (10 m lobster boats), field trip stops will be strongly weather dependent. On the trip south to Damariscove Island observe the steep, sometimes vertical sides of the river bank. Also note the absence of sediment cover on the adjacent terrestrial upland.

STOP 1: Damariscove Island.

From the larger vessels participants will be ferried ashore in small boats, either in Damariscove Harbor or the eastern cove, depending on swell conditions.

Damariscove Island was originally heavily forested like many of the surrounding islands, but was cleared in historic times by early settlers for firewood and pasture lands. It remains bare today. The coastal environments of Damariscove Island typify the sediment stripped, outer zone III of the estuarine model (Figure 3), consisting mostly of bare bedrock ledge in the intertidal region. Note the large boulders and blocks capping bare ledge exposed to open Gulf of Maine swells and storm waves, and limited boulder-cobble gravel pocket beaches in protected zones. Rockweed is dominant in the intertidal zone. The rest of the island is covered with thin soils and limited pockets of till, with no thick sedimentary sections. More sediment is found underwater, below about 10 m depth. Immediately adjacent to the island are paleovalleys cut into bedrock and Pleistocene glacial and glaciomarine sediments filled with more than 10 m of Holocene estuarine and marine muds. These sediments are best preserved below 50 m depth, with progressively less preservation in shallower water. Two former drainage pathways of the Damariscotta river can be traced, one immediately east of Damariscove, and also one east of Outer Heron Island (in the next chain of islands to the east).

TRANSIT 1:

After leaving Damariscove Island we will transit north up the river, past the towns of South Bristol and East Boothbay. The coast is composed primarily of rock ledge, with deep water close to shore. There is heavy forest on shore in most locations. Although coast land use is increasing for summer and year-round residences, the area is still primarily pristine. In any case, the rocky coast is stable, and the primary hazards to homeowners are wind and salt spray during storms.

Approximately 13 km north of Damariscove Island is the first of several bedrock sills that divide the Damariscotta River into a series of discrete basins. The most seaward sill at the southern end of Fort Island (Figure 4) rises to within 9 m of mean low water (MLW) and constricts to a width of 300 m. The basins on either side of the sill are more than twice the sill depth. The #10 nun buoy on the south side of the narrows has been observed rolled over on its side and completely submerged, implying a tidal current of at least 2.5 m/sec at maximum flow. North from Fort Island narrows to Wentworth Point is a gradual transition from the stripped outer zone III to the eroding central zone II (Figure 4).

STOP 2: I.C. Darling Center, Wentworth Point, Walpole, ME.
(LUNCH)

The Darling Center was a gift to the University of Maine from the late Ira C. Darling of Kennilworth, Illinois. In 1965, he donated this property of 130 acres, which includes over 2 km of riverfront property on Wentworth Point (the site of the dock) to the University with the stipulation that it be used as a marine laboratory. In addition, the University owns the next peninsula to the south, McGuire Point. That 27 acre property was given to the University of Maine by George Willett in 1975, and includes a picturesque five bedroom dwelling now used to house students and visiting investigators. These properties also abut Lowes Cove, a large tidal flat which has been the site of numerous detailed geological and biological studies by the lab's scientists. Facilities of the Darling Center include laboratory and office space, a flowing seawater laboratory and aquaculture facility, library, dormitories, conference and classrooms, diverse oceanographic sampling equipment, and a fleet of small boats, including the R/V *Miss Bess* and R/V *Lee*, our transportaion for today.

The attributes of the central eroding zone II are fully developed as we sail upriver to Wentworth Point. The central zone is charactertized by a predominance of mudflats which frequently front eroding bluffs, and small, discontinuous fringing marshes. On the Damariscotta River, thse flats, bluffs and marshes are best observed on numerous small- to medium-sized tributary bays on the margins of the main estuary. Examples includes Lowes Cove, just southeast of the dock, and Pleasant Cove, directly across the river to the southwest. The main river is primarily rock-framed well upstream. Above mean high water (MHW), the sediment cover is thicker, with thin basal till, washboard moraines, glaciomarine mud (the Presumpscot Fm.), and localized stratified drift.

The subtidal stratigraphy in the vicinity of the Darling Center dock is typical for the central Damariscotta River. The basement of Paleozoic

metasediments is frequently overlain by till. Conformably overlying the till is the Presumpscot Fm., also of highly variable thickness, but sometimes over 10 m. The lower units of the glaciomarine mud show rhythmic stratification in seismic profiles, which drapes underlying rock and till. This is interpreted as proximal, high sedimentation rate deposits produced by overflow and interflow processes. The upper units of the glaciomarine are ponded, and interpreted as distal density underflow deposits. Overlying the distinct unconformity is a blanket of Holocene mud, reworked from the underlying Presumpscot Fm. by tidal currents within the estuary and by slumps, waves, and currents which attack the marginal bluffs. A seismic signature which is common in the estuaries and nearshore shelf is a wipeout, or acoustically impenetrable zone. This is caused by biogenic methane gas in the sediments, as confirmed from cores. This gas is found primarily in muds thicker than 10 m, near the channel thalweg, and at the mouth of rivers, which have significant terrestrial organic supplies. The organic concentration, critical thickness and permeability of the mud, and possibly the stratigraphic position at the estuaries' turbidity maximum all control the presence of the gas.

Figure 7 shows some underwater vibracores from the central Damariscotta River estuary. Most penetrate to the Pleistocene Presumpscot Fm., which commonly causes refusal of our system. The Holocene sequences are quite uniform estuarine muds, with more or less detrital organic debris and shells. Two particularly important cores are DR-VC-5 and DR-VC-6. Vibracore 5 is 4.5 m long, taken 8.67 m below MHW, and penetrates the Presumpscot Fm. at 4.0 m in the core. It is remarkable in having a 3 m sequence of abundant *Crassostrea virginica*, which are now extinct within the the river. These may correlate in age with the oysters found in the middens at Damariscotta, discussed under Stop 5. Core DR-VC-6 is 7.54 m long, taken at 7.81 m below MHW, and penetrates the Presumpscot Fm. at 7.15 m. It is a long sequence of Holocene estuarine muds overlying a basal salt marsh peat. This peat has been radiocarbon dated at 6340 ± 55 yrs. B.P., our deepest and oldest reliable sea-level indicator.

TRANSIT 2:

Continuing the trip north up the estuary, we cross two more bedrock sills and their resultant basins. The first constriction and sill is 2.3 km north of the Darling Center dock at Merry Island narrows, across from Kelsey Point (Figure 4). The second is only 1.5 km farther, at Glidden Ledges. The basin formed between the two constrictions is the smallest along the river. North of Glidden Ledges the conspicuously eroding bluffs are mostly washboard moraines (Smith, 1982) that can be traced on both sides of the river. These moraines were deposited during deglaciation and subsequently have been eroded during the Holocene transgression. Between the moraine are lower bluffs composed primarily of Presumpscot Fm.

Just south of Dodge Upper Cove on the west bank of the river is a striking example of high-tide pruning of terrestrial vegetation. The lowermost branches of trees mark a distinct horizontal datum that approximates spring high tide. This phenomenon is best seen on vertical rock shorelines with overhanging conifers.

STOP 3: Dodge Upper Cove Peninsula

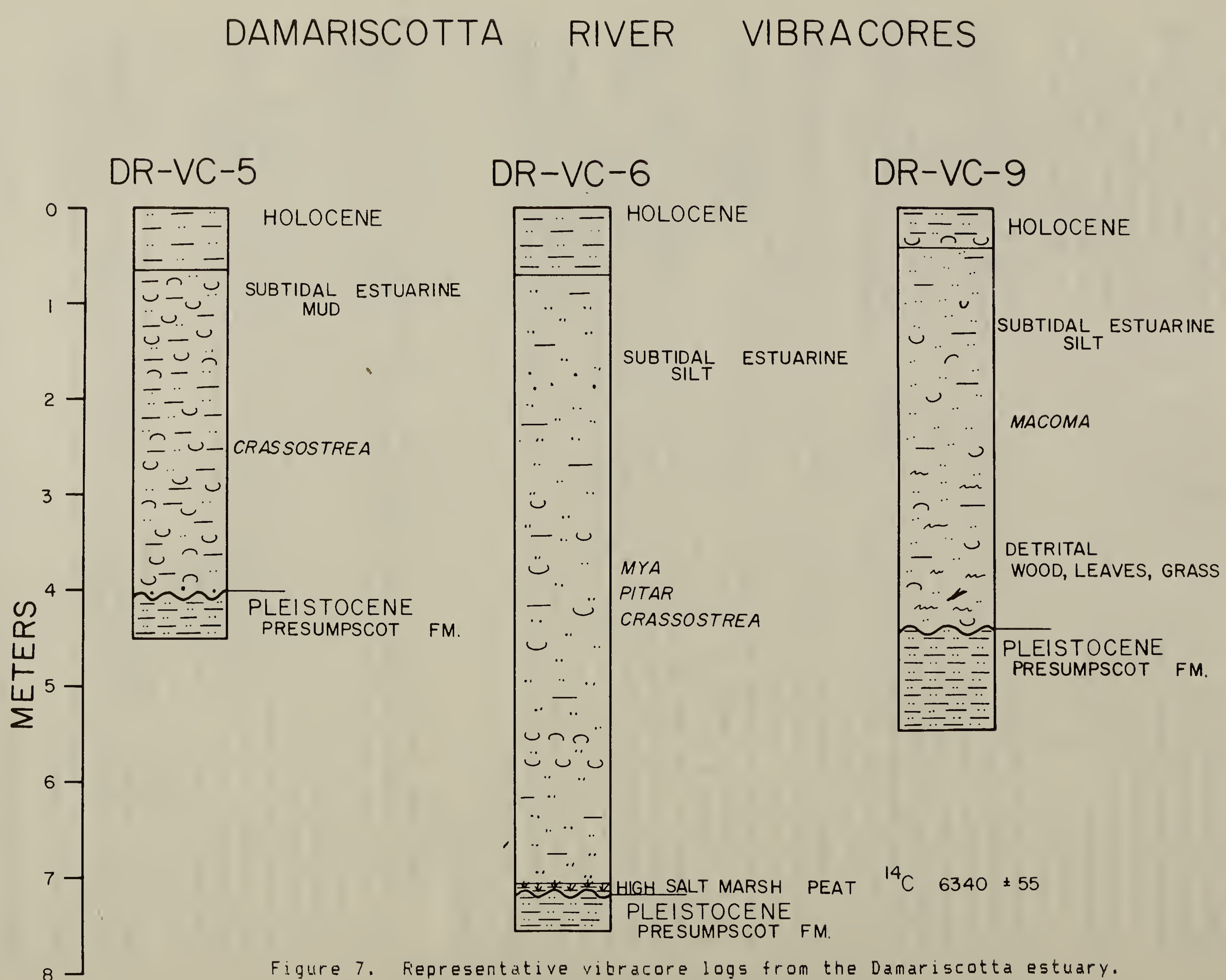

Figure 7. Representative vibracore logs from the Damariscotta estuary. Vibracore locations are shown in Figure 4.

Small boats will again be used to transfer the participants from the lobster boats to the eroding peninsula in Dodge Upper Cove. The eroding face on the south side of the peninsula consists of a basal diamicton overlain by interbedded mud and sand, all capped by a thin soil zone. We interpret the peninsula as an east-west trending minor (washboard) moraine. These washboards are strongly evident in airphotos in this section of the coast. The trend of this moraine can be traced in seismic data and onto the eastern bank of the river.

Toward the base of the peninsula on the south-facing gravel beach is evidence of strongly episodic sediment transport. A small recurved spit with several recurves is present on the high intertidal beach. The orientation indicates transport from the east. The episodic nature is suggested by a moderately dense marsh grass stand that covers the form during the summer. Other geomorphic evidence for direction of dominant winds and waves is provided by a comparison of the north and south faces of the peninsula. Toward the north the marsh fringing the tidal flat is narrow and scalloped along the margins. In addition, the high-tide line is marked by driftwood and the mudflat is lttered with boulders. Conversely, on the southern side of the peninsula, the marsh is thick and broader, while the high tide shoreline is marked with less debris and fewer boulders are found on the flat. It appears that the dominant storm event is from the northeast, which is consistent with local weather patterns.

Another obvious feature in Dodge Upper Cove are the red brick beaches. There are three sites in the cove composed of broken and deformed bricks. The brick-making industry on the Damariscotta thrived during the eighteenth and nineteenth centuries, peaking between 1860 and 1870 (H.W. Castner, unpub. manuscript). At least 30 brickyards were known along the river. The last brickyard along the river was operated well into the 1900's by Webster Kelsey of South Bristol. (His grandson Thurlow Kelsey is an employee of the Darling Center, and is an excellent source of local lore). Brickyards were prevalent here due to the riverbank exposures of Presumpscot Fm., excellent docking sites, and abundant firewood. Brickmaking was a summer activity, and consumed an average of 200 cords of firewood a season. Most bricks were shipped by schooner to better markets. The red brick beaches are a consequence of the discard of broken or deformed specimens. This concentration of brickyards is apparently unique along the Maine coast, possibly because of the finer-grained nature of the glaciomarine mud locally, which is homogeneous with few dropstones.

TRANSIT 3:

Proceeding north from Dodge Upper Cove the transition from zone II to the accumulating inner zone I occurs. Starting at Perkins Point (Figure 4), just north of the last stop, extensive intertidal and subtidal mudflats begin to rim both sides of the estuary. The typical inner zone I is characterized by marsh backed by stable bluffs, and mudflats backed by eroding bluffs. The inner zone I of the Damariscotta River is unusual in that no large marshes are found anywhere along the estuary. Fringing marshes are present, especially in protected coves. Many of the bluffs in the inner zone are stable, even though they are fronted by mudflats, exposed to waves and ice erosion. It is unclear why there are no broad marshes in this system, despite apparently amenable

locations. It may be a lack of the numerous tributary valleys, which allow the Sheepscot River estuary to maintain abundant broad marshes.

STOP 4: Linnet Cove

If tide level and time permit, we will illustrate a typical inner zone I environment at Linnet Cove. This area is one of our Sea Grant project stations being used to monitor bluff erosion and flat accretion. The significant differences between zone II and zone I are the unstable versus stable nature of the bluffs, and the increased volume of intertidal and shallow subtidal sediment accumulation. The bluff at Linnet Cove is more stable than the bluff visited at the last stop, as can be noted from the complete vegetation cover. Slumping does occur, however, as can be seen in a recently active slump scar. Slumping is a common mechanism both in zones I and II, but is more important in zone I due to the lesser effects of direct wave and current attack. Slumping is thought to provide the majority of sediment to the Holocene systems of this zone (Belknap and Kelley, unpub. Sea Grant Proposal, 1986). Kelley's trip B-8 in Casco Bay this year shows this process in more detail.

On the intertidal beach in Linnet Cove is an array of stakes and buried steel plates which we are using to monitor horizontal movement of the bluff and vertical accretion or erosion on the tidal flats, respectively. The stakes are surveyed in, and then measurements with a tape to the bluff are taken several times a year in order to determine rates of change. The buried plates are probed at the same time, to measure depth of sediment. Five of these arrays are located in the Damariscotta Estuary, and 18 others are located elsewhere along the Maine coast (in Casco Bay and Machias Bay at present). The purpose of this network is not only to determine trends and rates of change of bluffs and flats, but also to serve as a baseline data set. These data will be particularly usefull in assesing the effects of a postulated increase in tidal range across the Gulf of Maine if the proposed Canadian tidal power projects in the Bay of Fundy are completed (Greenberg, 1979; Larsen and Topinka, 1984).

TRANSIT 4:

North of Linnet Cove greater than 75% of the river basin is either intertidal or shallow subtidal (<1 m depth at MLW). In fact, any boats moored in this area must remain in the channel or they will settle onto the flats. The broad basin surrounded by the towns of Damariscotta and Newcastle has been affected somewhat by growth in land use, changing from a ship-building and brick-making area to the present retirement village and shopping center for the Bristol Peninsula. Accumulation of mud and encroachment of marsh over the former ship-building ways is evident, and attested to by local historians. The rapid shoaling of this basin since colonial times has not yet been confirmed geologically, but coring and comparative map and chart studies are underway.

We will disembark from the vessels at the Damariscotta town dock.

STOP 5: Damariscotta River: Glidden Point Shell Middens

After leaving the boats at the Damariscotta town dock, we will reboard the vans. If there is enough time, we will visit the Damariscotta shell middens. Leaving the parking lot, turn left and cross the bridge to Newcastle, turn right on the Mills Road to the Route 1 entrance ramp (0.8 km; 0.5 mi), turn north, proceed 1 km (0.6 mi) to the Route 1 bridge, park on the west approaches.

The Damariscotta shell middens are composed almost exclusively of *Crassostrea virginica* shells, disarticulated and mounded into discrete layers which coalesce into the major middens on the north, east, and west sides of this bend in the Damariscotta River. They have been variously estimated to have contained between 5 and 45 million cubic meters of shells (Castner, 1950) before extensive mining for agricultural lime in the late 1800's. The middens have paleoecological as well as archaeological significance because there are no viable oyster populations in the river today, and only limited isolated occurrences in the Sheepscot River nearby. Even oyster culturing has not been successful. The oysters in the midden, however, are the largest known in the world, some over 35 cm long. Shell middens elsewhere in the river and along the Maine coast are made up primarily of *Mya arenaria*, and are much smaller.

Richard Goldthwait (1935) used the Damariscotta shell heaps as a constraint on local relative sea level. A series of bedrock sills between Damariscotta-Newcastle and this location limit tidal range. One sill, Johnny Orr, is only 1 m below MHW. Thus sea level could only have risen that far since the midden was produced. Myers (1965) dated shells which suggested that the deposit was 2000 years old. Sanger and Belknap (1986 in press) have recently re-examined the sea-level implications of these middens. Radiocarbon dates from the base of the midden range from 1600 to 2300 yrs. B.P. Artifacts at the top of the midden suggest abandonment at least 700 years B.P. The later point is important, because local lore claims that sawdust smothered the oysters; in fact, the major productivity was over long before colonial times. A similar result obtains for increased runoff from colonial deforestation and agriculture. Newell (1983) studied the biology of the oysters, and found that all factors in the Damariscotta River today are conducive to oyster growth and reproduction (salinity, temperature, turbidity, substrate) except predator load. Oysters can tolerate as little as 6 ppt (parts per thousand) salinity, with 10-30 ppt most acceptable. Currently, salinity above Johnny Orr ranges from 15 ppt in spring to 29 ppt in summer and fall. The predator *Urosalpinx sp.* (oyster drill) cannot survive salinity less than 20 ppt (the "drill line"). Thus, it is probable that oysters colonized an optimal location in the Glidden Point bend as sea level overtopped the bedrock sill 2300 yrs. B.P. Indians used the resource as a primary food source for about a millennium. Continuing sea-level rise allowed the predator *Urosalpinx* to intrude, further decimating the oysters. Both man and oyster drill may have contibuted to the oysters decline, but their relative importance is unknown. Oysters are found in subcropping defunct reefs in Salt Bay above this point, and are found in vibracores in the lower half of the river. It is likely that the ecological changes described above were repeated more than once.

This is the final stop of our itinerary. Reboard the vans for the return trip to the Darling Center to pick up the cars.

TRIP C-1

GLACIAL GEOLOGY OF THE WHITE MOUNTAIN FOOTHILLS, SOUTHWESTERN MAINE

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INTRODUCTION

The purpose of this field trip is to examine some of the glacial features in the area between Fryeburg and Bethel in southwestern Maine. The Maine Geological Survey is mapping the surficial geology of this area, but the Quaternary stratigraphy and history have yet to be worked out in detail. Therefore our trip will highlight some of the more interesting and problematic glacial deposits, rather than attempting a comprehensive review of the regional stratigraphy.

Previous investigations of the glacial geology of the Fryeburg-Bethel area were carried out by Stone (1899), Leavitt and Perkins (1935), and Prescott (1979; 1980a; 1980b). Thompson has mapped the boundaries of sand and gravel aquifers (Williams et al., in prep.), and is compiling detailed 1:24,000 surficial quadrangle maps for this part of the state. W. R. Holland (in prep.) has mapped the surficial geology of nearby quadrangles between Fryeburg and the Ossipee River. (See Trip B-1, this volume.)

Much of the field trip area is a rugged and scenic terrain that includes part of the White Mountains and their foothills. The mountains just north of Kezar Lake separate north-draining tributaries of the Androscoggin River from the Saco River basin to the south. The flow of glacial ice has caused many hills to have asymmetric shapes, with sheer bedrock cliffs resulting from plucking on their southeast sides and more gentle stoss slopes to the northwest. Drumlinoid hills (commonly rock-cored) and striations on bedrock provide additional evidence of predominantly south-southeastward glacial flow. The major valleys--especially those of the Saco, Androscoggin, and Cold Rivers--contain thick accumulations of water-laid glacial and postglacial sediments. Kames and eskers are seen in the Androscoggin Valley, but ice-contact and glaciolacustrine sediments in the Saco Valley are concealed beneath a broad alluvial plain.

It is hoped that ongoing field work in southern Oxford county will clarify the following aspects of the area's glacial history: (1) the till stratigraphy and chronology of glaciations that it represents; (2) the pattern and style of deglaciation, especially the relative importance of active-ice retreat vs. regional stagnation in generating glaciofluvial and glaciolacustrine sediments; and (3) the glacial lake history. The field trip stops will provide a focus for discussion of these topics.

DESCRIPTION OF STOPS

STOP 1: Keewaydin Lake Till Section

Three units are exposed in this section:

(1) The upper unit (just below the ground surface) consists of about 1 m of sandy, stony glacial diamicton. This material is probably an ablation till deposited by the late Wisconsinan ice sheet, although it may have been modified by postglacial colluviation. A thicker exposure of the same till unit can be seen a short distance east of here, on the south side of Route 5 adjacent to Keewaydin lake.

(2) The middle unit is about 1 m thick, and consists partly of disrupted slabs and irregular fragments of the till seen in the lowest unit. These till fragments are interlayered with thin, discontinuous, deformed lenses of sand, causing the unit to have a crude stratification. Scattered stones are also present. This unit may be a mixture of the upper and lower tills that resulted from erosion of the lower till by the last ice sheet. Pessl and Schafer (1968) described a similar mixed zone along the erosional contact between two tills in Connecticut, and the author has seen this type of contact at other multiple-till localities in southern Maine. At the Keewaydin Lake section, sand dikes extend from the mixed zone down into the underlying till unit.

(3) The lower unit has an exposed thickness of at least 12 m, and consists of a silty, olive-gray till. This till has a sharp contact with the middle unit. It is finer-grained than the upper till and has a blocky structure resulting from jointing. A dark-brown Fe/Mn-oxide stain coats joint surfaces in the upper part of this unit, as well as in the till inclusions in the mixed zone described above. The compactness, structure, and fine-grained texture of this till indicate that it is probably a lodgment facies.

The upper till at Stop 1 is thin and poorly exposed, but the characteristics of the two tills and their contact relationships invite comparison with the two-till stratigraphy of New Hampshire described by Koteff and Pessl (1985). The surface till of New Hampshire and southern New England is typically non-oxidized (except in the zone of modern soil formation) and is assumed to have been deposited by the late Wisconsinan ice sheet. The age of the lower till is uncertain; it may be early Wisconsinan or pre-Wisconsinan (Thompson and Borns, 1985). Both the oxide staining and the matrix oxidation seen in the upper part of the lower till suggest a weathering interval before deposition of the upper till. This inference has been supported by the few comparative mineralogical studies that have been done on the two tills.

STOP 2: Bryant Hill Pit

This is one of the most complex exposures of Pleistocene sediments in the field trip area. A long northeast-trending pit face, up to about 25 m high, reveals a thick deposit of stratified sand and gravel with interbedded diamicton lenses. The upper part of this deposit is cut by thrust faults in the northeastern part of the pit, and the sand and gravel is overlain by one

or more till units. Striations on two nearby bedrock outcrops along Route 5 indicate local ice-flow directions of 184° and 195° (SSW).

The stratified unit, which forms much of the section, has an exposed thickness of up to about 23 m. It ranges from well-sorted silt and sand to boulder gravel, but consists chiefly of thin beds of compact, poorly-sorted pebbly sand to cobble gravel. The bedding dips generally southward at 10-15°, and tends to be laterally continuous over distances of several meters or more. A few large boulders are scattered through the unit, and there are conformable lenses of olive-gray, silty to sandy diamicton. The origin of this stratified unit is not clear. Did it form as a subaerial outwash fan; as a delta or subaqueous fan in an ice-marginal lake; or even in a cavity beneath the ice? An ice-contact environment seems likely because of the angularity and poor sorting of the gravel component, the presence of striated stones and large boulders, and especially the occurrence of the diamicton lenses. The latter are probably flowtills which avalanched from nearby glacial ice. The evenly dipping sand and gravel layers resemble delta foreset beds at first glance, but on closer inspection they lack the ripple-drift cross-stratification and graded bedding that occur in some delta foresets. The stratified unit is situated in the lee of a bedrock knob (which outcrops on the summit and northeast flank of Bryant Hill), so it may have been deposited in a subglacial cavity that developed as ice flowed over the hill. However, the considerable thickness and extent of the unit cast doubt on the latter theory.

In the southwestern part of the pit, the sand and gravel unit is overlain by about 6 m of sandy, stony till. This till has a weak fissility, and contains thin, discontinuous beds of silt and sand. The manner in which these stratified lenses drape over stones suggests that the till is a basal melt-out deposit.

In the central and northeastern part of the pit face, the upper part of the stratified unit is offset by several low-angle thrust faults (Figure 1). These faults have not yet been studied closely because of their location high on the pit face, but they appear to displace the sand and gravel beds toward the south. They developed as ice overrode the north end of the deposit, possibly as a consequence of the same flow that engraved the SSW-trending striations on nearby outcrops. Slickensides were found on the bedding planes of silt layers within the sand and gravel at the northeast end of the pit, and may have resulted from the same deformational event as the major faults.

The fault zone is overlain by a till unit that is 1-2 m thick, more bouldery toward the northeast corner of the pit, and contains highly deformed sand lenses. It is not clear whether this till was deposited in the same manner as the till at the other end of the pit, nor whether they were deposited at the same time. One of the thrust faults, seen near the top of the section in the central part of the pit face, seems to separate an overlying pebbly till(?) from the thicker, bouldery till unit to the southwest.

STOP 3: Hatch Hill Pit

Hatch Hill exhibits a glacially streamlined shape on the Center Lovell quadrangle, as do other NNW-SSE trending hills in the immediate area.

Figure 1

Thrust faults in upper part of stratified sand-and-gravel unit at Bryant Hill pit (Stop 2). View is to the west, in the northern part of the high pit face.

Figure 3

View southwest across the Androscoggin Valley, showing high moraine ridge projecting eastward from Stock Farm Mountain (center to left-center) and juncture of the moraine system with Hark Hill (shaded area in foreground).

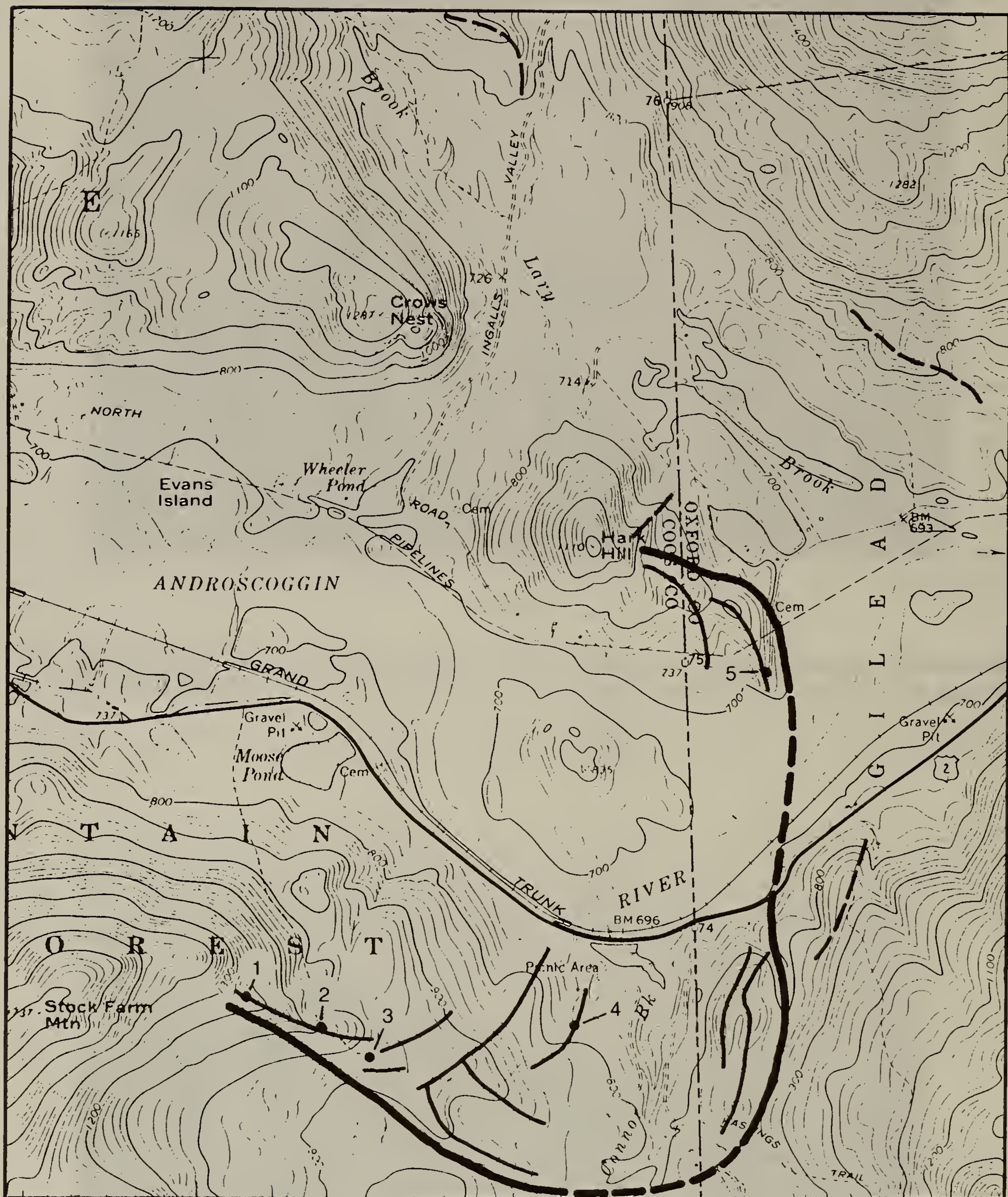

FIGURE 2. Southeast part of the Shelburne 1:24,000 quadrangle, showing outline of the principal part of the Androscoggin Moraine system. Stop 6 is located on the north end of this moraine cluster, just southeast of Hark Hill. Narrow lines indicate the crests of moraine ridges. Numbers designate locations of test pits.

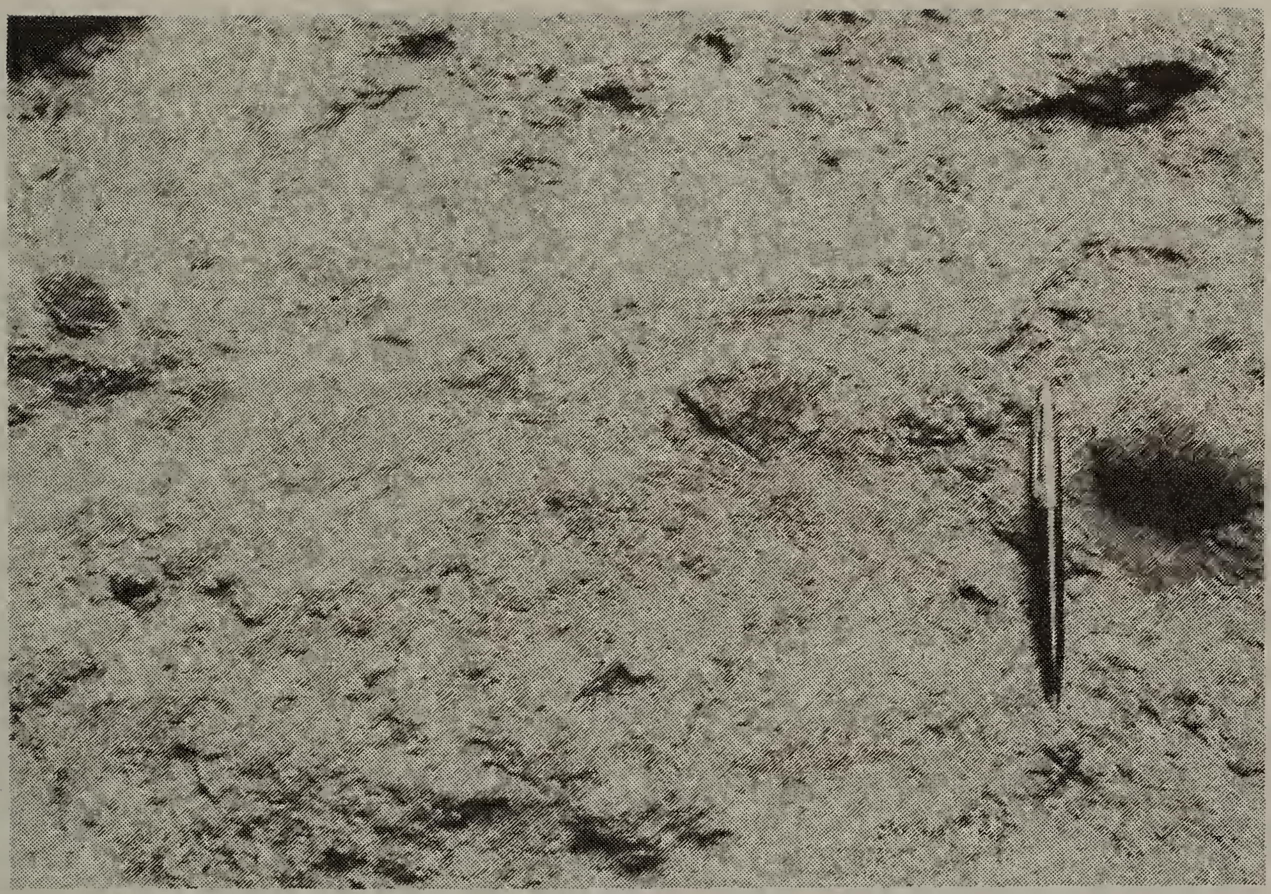

Figure 4

Close-up of deformed silt-sand laminae in sandy, stony glacial diamicton (till) in Test Pit 3, Androscoggin Moraine. Pen is 13 cm long.

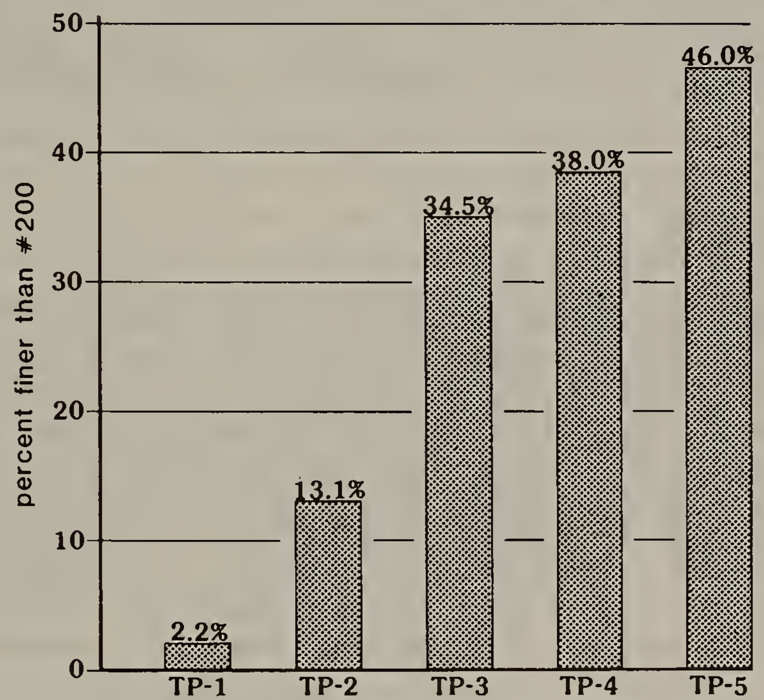

FIGURE 5. Graph showing weight percentage of silt and clay fraction (percentage passing #200 sieve) in bulk samples from Test Pits 1 - 5.

Although it has this drumlin-like shape on the map, the surface of the hill is actually more uneven than an ideal drumlin; and bedrock outcrops in the small field on the summit.

The exposure in the pit seen at Stop 3 is located on the southeast end of Hatch Hill. Like the previous stop, it shows a glacial diamicton overlying stratified drift in the distal part of the rock-cored hill. Having seen this relationship at two closely spaced localities, one wonders how many other sand and gravel deposits may lie concealed beneath tills and associated glacial diamictons in other such hills in this part of Maine.

In detail, the Hatch Hill deposit differs in several respects from the Bryant Hill locality. The lower unit seen here consists of well-stratified sand and pebbly sand, which dips generally south-southeast and is cut by normal faults. The sand displays foreset-type bedding that appears to have been deposited in a lake or subglacial cavity.

The sand unit is overlain by about 2-8 m of complex glacial diamicton. This diamicton is stratified due to the presence of abundant sand and silt layers that are interbedded with the till-like sediment. Large boulders (some of which are well striated) occur within this unit and on the ground surface adjacent to the pit. Apart from the prominent stratification, at least two other features distinguish the diamicton unit from tills seen at many other localities in southern Maine. First, it contains many clasts of laminated silt and sand, resulting from disruption of these strata by a debris flow, or ice-shove. These inclusions are best seen near the top of the section on the northeast side of the pit. Second, the diamicton also contains odd structures consisting of vertical fissures or chimney-like cylindrical openings that were somehow eroded and then filled with sand and gravel.

The same question arises here that was posed at the previous locality: is the sandy unit a glacial-lake delta that was buried by the diamicton as a consequence of a later glacial advance, or was the sand deposited in a subglacial cavity (in which case the diamicton may have simply melted out of the overlying ice)?

STOP 4: Bradley Brook Delta

This delta was deposited where Bradley Brook presently empties into the Cold River valley. The contact between the topset and foreset beds is at an elevation of about 500 ft, and marks the position of the water level to which the delta was graded. This deposit may have been built by glacial meltwater streams flowing down the Bradley Brook valley from the high mountains to the west. However, at least two questions arise here. One is the location of the dam that impounded the lake. Test-hole data from the center of the Cold River valley (just east of the delta) reveal the presence of lake-bottom sediments beneath the flood-plain alluvium. Therefore, the lake extended across the valley and was not just a minor ice-contact pond confined to the immediate area of the delta. At present there is no barrier to contain a lake in the south-draining valley, so it is inferred to have been dammed by a temporary plug of stagnant ice and/or glacial sediments. A logical site for the former dam is the narrow part of the valley just southeast of here, where there is a large sand and gravel deposit extending to an elevation of 500-520 ft.

The other problem is the source of the sediment that comprises the delta. This deposit is not large, but the drainage basin of Bradley Brook seems to have a very small area from which to derive the quantity of sand and gravel seen here. Also, it is not certain to what extent the sediment came from ice masses in the hills, if in fact the valley bottom was largely free of ice and already contained a lake. Perhaps the delta was derived partly from erosion of the steep mountainsides soon after they were deglaciated. Any ice that did linger in the headwaters of Bradley Brook would have been essentially stagnant because of its position in the lee of the Baldface Range.

STOP 5: Evans Notch Meltwater Channel and Alluvial Fan

The parking lot at the head of East Royce Trail is situated in the highest part of Evans Notch, precisely upon the drainage divide between the Androscoggin and Saco River basins. Just northeast of here (along the upper portion of Evans Brook), there is a low, swampy area marking the path of a glacial meltwater channel through the notch. Meltwater draining through this channel was at least partly responsible for carving the deep, spectacular part of the notch to the south. Ice-contact sand and gravel (seen along Route 113) was graded to the notch when glacial ice still filled the Evans Brook valley at least to the level of the divide at Stop 5. The concept of generally northward recession of the ice margin in this area is further supported by other meltwater channels and associated deposits at lower elevations between Evans Notch and the Androscoggin River.

At Stop 5 the Evans Notch channel is not apparent because it has been buried by the postglacial alluvial fan on which the parking lot was constructed. This fan consists of bouldery alluvium deposited by the brook that descends the steep hillside to the west of here. The brook divides on the surface of the fan, with distributary channels extending into both the Androscoggin and Saco basins. During a recent visit when conditions were rather dry, the author noted that surface drainage was only to the north, and the water disappeared into the fan before reaching its outer edge. It is interesting to observe that this sizable deposit has been constructed by such a small stream in Holocene time. As a matter of speculation, it is likely that much of the fan was built during major floods of infrequent occurrence, and the rate of aggradation may have varied over postglacial time. This stop also raises questions as to the magnitude of Holocene stream sedimentation in the White Mountains as a whole, which may be greater than generally realized. For example, a stream-terrace along part of nearby Evans Brook (see Road Log) may be dissected Holocene alluvium rather than glacial outwash. Large alluvial fans can be seen in the upper Androscoggin River basin, as in the Lary Brook valley (northwest of Stop 6, Figure 2) and in the vicinity of Gorham, New Hampshire.

STOP 6: The Androscoggin Moraine

The Androscoggin Moraine is a glacial end-moraine system that is located in the vicinity of the Maine-New Hampshire border in the Androscoggin River valley. Parts of this moraine system were described by Stone (1880), and the segment seen at Stop 6 was illustrated in his USGS Monograph entitled "The

Glacial Gravel of Maine" (Stone, 1899). Leavitt and Perkins (1935) disagreed with Stone's identification of this ridge as a moraine, but they did not present convincing evidence to the contrary. Thompson (1983) agreed with Stone's interpretation, and reported the discovery of another prominent moraine ridge extending from Stock Farm Mountain on the south side of the valley.

A detailed investigation of the moraine system is currently in progress, with the objective of confirming that it was deposited by a tongue of ice extending down the Androscoggin Valley. Other goals are to determine the composition of the moraine, its provenance, and its regional significance in the deglaciation history of western Maine and the White Mountains. The Androscoggin Moraine is an unusual and significant feature because moraines are rare above the zone of late-glacial marine submergence in Maine, and it is the most clearly defined of the very few moraines in the White Mountain region. It indicates that active ice persisted in the upper Androscoggin Valley at a time when the mountains must have emerged from the thinning late Wisconsinan ice sheet. The Androscoggin ice tongue is also believed to have deposited the Success Moraine on the northwest flank of the Mahoosuc Range (Gerath et al., 1985) and possibly the till that overlies lake sediments in the Peabody River valley south of Gorham (Gosselin, 1971). The age of the Androscoggin Moraine is uncertain, and depends on whether the moraine was deposited by the Laurentide Ice Sheet or the Appalachian ice mass that was separated from the Laurentide by opening of the marine embayment in the St. Lawrence Lowland about 13,000 years ago. It is hoped that studies of till provenance and striation patterns (now in progress) will help solve this problem by determining whether the ice lobe was restricted to the Androscoggin Valley, or if part of the moraine was derived from ice that was thick enough to extend eastward from the Connecticut River valley.

Results obtained to date have essentially confirmed that the high ridges on both sides of the valley (Figure 2) are truly moraines, and not just a deceptively oriented group of drift tails and other deposits. Correlation of moraine segments across the valley is difficult, but it is evident that the Androscoggin Moraine complex represents several closely spaced ice-margin positions.

Numerous large boulders (1-8 m in diameter) are strewn along the ridges shown in Figure 2. The higher moraine segments are very steep-sided, and the one that projects from Stock Farm Mountain (Figure 3) locally stands as much as 30 m above the adjacent terrain. The moraine crests rise in elevation from about 720 ft (where breached by the Androscoggin River) to 900 ft on Hark Hill and 1250 ft on Stock Farm Mountain. Bedrock outcrops extensively on these hills, and on low hillsides just upvalley from the moraine. However, extensive search has not revealed any outcrops along the moraine ridges. The local bedrock is gneissic Littleton Formation which is intruded by quartz diorite (Billings and Fowler-Billings, 1975). Boulders of these rock types are very conspicuous in the moraine system.

Exposures of the sediments comprising the Androscoggin Moraine are rare, and they are limited to a few shallow cuts along the logging road that follows the crest of the moraine ridge on Stock Farm Mountain. During the early summer of 1985, five test pits were dug with the aid of a backhoe in order to determine the composition of the moraine and collect sediment and stone

samples. Four pits were dug along the logging road on the east side of Stock Farm Mountain, and one on the easterly moraine segment that extends from Hark Hill (Figure 2). A variety of glacial sediments were encountered in these pits, as one would expect to find in deposits formed in an ice-marginal environment. Test pits 1-4 exposed various facies of tills and flowtills with some contorted lenses of sand, silt, and gravel (Figure 4). Test Pit 5, located in the lowest part of the moraine, exposed a water-laid diamicton consisting of interbedded silt, fine sand, and flowtill(?), with angular stones present in all of these units. Bulk sediment samples from the five pits are progressively finer grained with decreasing elevation above the valley floor (Figure 5). This trend is the result of two factors: the greater removal (by meltwater) of silt and clay from the sandy-matrix diamictons at higher elevations on the Stock Farm Mountain moraine segment, and the presence of numerous silty laminae indicative of local meltwater ponding in the vicinity of Test Pit 5.

Preliminary results of stonecounts (on 100 stones from each of four pits) indicate that local rock types predominate in the moraine system, but there is a marked difference between the two sides of the valley. Exotic fine-grained igneous dike rocks are much more common in the Hark Hill segment of the moraine. These erratics probably were derived from dike swarms in the Gorham-Berlin area, which have been described in detail by Billings and Fowler-Billings (1975).

ACKNOWLEDGMENTS

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ITINERARY

Assembly point: Rest area on south side of Route 117, just west of Norway village (at south end of Pennesseewassee Lake). Virtually all of the itinerary is covered by Map 10 in the Maine Atlas. Topographic map coverage for the field trip stops is provided by the East Stoneham, Center Lovell, Speckled Mountain, and Shelburne 7.5-minute quads, and the North Conway 15-minute quad.

Total Mileage	Mileage Between Stops	
0.00	0.00	Leave rest area and go west on Route 117.
0.45	0.45	Jct. with Route 118. Proceed straight ahead (W) on Route 118.
11.30	10.85	Jct. with Route 35 in North Waterford (end of Route 118). Go straight ahead (W) on Route 35.
12.30	1.00	Jct. with Route 5 in Lynchville (note famous road sign showing distances to Peru, Mexico, etc.). Continue straight ahead (W) on Route 5.
15.60	3.30	Stop 1: Keewaydin Lake till section (East Stoneham 1:24,000 quadrangle). Park at edge of Route 5 and climb up to excavated bank on S side of road.
18.70	3.10	Stop 2: Bryant Hill pit (Center Lovell 1:24,000 quadrangle). Park at edge of Route 5 and walk into large pit to W of road.
23.80	5.10	Turn R onto West Lovell Road (sign points to Kezar Lake).
23.90	0.10	Turn R (small sign designates this road as "Grovers Lane").
24.05	0.15	Turn L onto dirt road and drive to pit.
24.25	0.20	Stop 3: Hatch Hill pit (Center Lovell 1:24,000 quadrangle).
24.50	0.25	Upon return to Grovers Lane, turn R.
24.60	0.10	Turn L onto West Lovell Road and return to Route 5.
24.70	0.10	Turn R and proceed S on Route 5.

25.05	0.35	Note dissected glacial-lake sand deposits to R (under golf course).
27.80	2.75	Turn R onto Shave Hill Road (sign says "to Route 113").
28.75	0.95	Keep L at fork in road.
29.30	0.55	Enter Saco River flood plain.
30.30	1.00	Turn R onto Union Hill Road.
30.70	0.40	Keep L at road jct.
32.30	1.60	Turn L (onto "New Road" in Maine Atlas).
33.45	1.15	Jct. with Route 113 at Stow. Go straight ahead on Route 113.
33.60	0.15	Keep to R at jct., proceeding N on Route 113.
36.70	3.10	Stop 4: Bradley Brook delta (North Conway 1:62,500 quadrangle). Turn R into gravel pit on E side of Route 113. Stay on hard-packed road--avoid soft sand.
36.75	0.05	Exit pit via N entrance. Turn R and continue N on Route 113.
41.35	4.60	Turn L on road to "The Basin" (White Mtn. National Forest picnic area).
41.90	0.55	Lunch stop at The Basin.
42.65	0.75	Return to Route 113. Turn L and continue N on Route 113.
45.85	3.20	Stop 5: Evans Notch meltwater channel and alluvial fan (Speckled Mountain 1:24,000 quadrangle). Turn L into parking lot for East Royce Trail.
45.90	0.05	Turn L out of parking lot. Continue N on Route 113.
46.05	0.15	Note Evans Notch spillway channel on L.
48.70- 48.80	2.65- 2.75	Note stream terrace in woods to E of road. The origin of this deposit is uncertain. It may be glacial outwash or Holocene alluvium.
53.50	4.70	Turn R onto U.S. Route 2 at Gilead.
53.85	0.35	Turn L. Cross RR track and the Androscoggin River.
54.20	0.35	Turn L onto North Road and continue W along the Androscoggin Valley.

56.50	2.30	Stop 6: Androscoggin Moraine (Shelburne 1:24,000 quadrangle). Park where pipeline crosses North Road. Walk along pipeline to ridge crest just W of road.
56.80	0.30	Cross second moraine ridge.
57.10	0.30	Note roche moutonnee outcrop in pipeline clearing to N of road. Shape of outcrop indicates ice flow to east, parallel to the valley.
57.20	0.10	View of Stock Farm Mtn. on other side of valley (to S). Note the high part of the Androscoggin Moraine projecting E from the mountainside.
59.30	2.10	Turn L at jct. and cross Androscoggin River at Shelburne.
60.20	0.90	Turn L onto U.S. Route 2.
63.95	3.75	Park along U.S. Route 2, a short distance E of state line. Looking back up the valley offers a good view of the Androscoggin Moraine, which extends E from Stock Farm Mtn. View is best in late fall or winter, with low sun angle and no leaves on trees.

END OF TRIP

Trip C-2

A GEOLOGIC TRAVERSE WITHIN THE EASTERN EDGE OF THE MEDIAL NEW ENGLAND TERRANE

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The Medial New England terrane underlies much of Maine, and it contacts a more easterly terrane (Avalon or Castine-Ellsworth block of Stewart and Wones, 1974) approximately along the coast. The Medial New England terrane consists of a probable composite basement and cover sequences of Late Precambrian through Lower Devonian age. Silurian-Devonian rocks are extensively exposed in the Merrimack is Synclinorium, and older rocks are exposed as inliers and along the east edge of the Medial New England terrane. Most of the structural, the plutonic and the metamorphic features in Maine resulted from the juxtaposition of the Medial New England and the more easterly terrane in Devonian time. This field excursion provides representative exposures of the rocks, some of the structural features and metamorphic character of the eastern edge of the Medial New England terrane.

Trip Cancelled 7-17-86

BEDROCK GEOLOGY OF THE NEWFIELD AND BERWICK QUADRANGLES, SOUTHERN MAINE

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INTRODUCTION

This trip will emphasize the stratigraphy and structure of Siluro-Devonian metasedimentary rocks found in the southern part of the Newfield (15') Quadrangle and the northern part of the Berwick (15') Quadrangle. Bedrock mapping in the 1960's and 1970's by Gilman (1977) and by Hussey (1968) resulted in assigning the metasedimentary rocks to the Rindgemere Formation, believed at the time to be correlative to the Lower Devonian Littleton Formation in New Hampshire. Subsequent study of new road cuts along the Spaulding Turnpike north of Rochester, N.H. by Eusden (1984) led to a proposed correlation of those rocks with the established Siluro-Devonian sequence of the Rangeley area (Moench, 1971) which Hatch, Moench, and Lyons (1983) traced to the southwest into central and eastern New Hampshire. We will examine alternative extensions of the sequence from the Spaulding Turnpike into the southern part of the Newfield Quadrangle as proposed by Eusden (1984, 1986), Hussey (1985) and by my recent mapping.

REGIONAL GEOLOGY

Sillimanite zone metasedimentary rocks on the eastern limb of the Kearsarge-Central Maine Synclinorium (KCMS; Lyons and others, 1982) are extensively exposed throughout the Newfield Quadrangle as well as the Kezar Falls area to the north (Gilman, 1977) (figure 1). These rocks consist of well-bedded to non-bedded, non-rusty, pelitic schists and migmatites interlayered with sulfidic, rusty weathering schists, and a distinctive thinly-bedded, grey-green, calc-silicate granulite with associated fine-grained granular biotite-quartz-feldspar schist. A non-migmatized, well-bedded schitose quartzite found south of Acton shows the best potential for along-strike correlation with the sequence along the Spaulding Turnpike.

Structurally, the metasedimentary rocks are characterized by a gently dipping bedding plane schistosity. Dips rarely exceed 30 degrees and horizontal attitudes are not uncommon. Minor folds seen at a single outcrop almost always fold the schistosity indicating multiple deformations. Where facing directions can be determined from graded bedding, it is com-

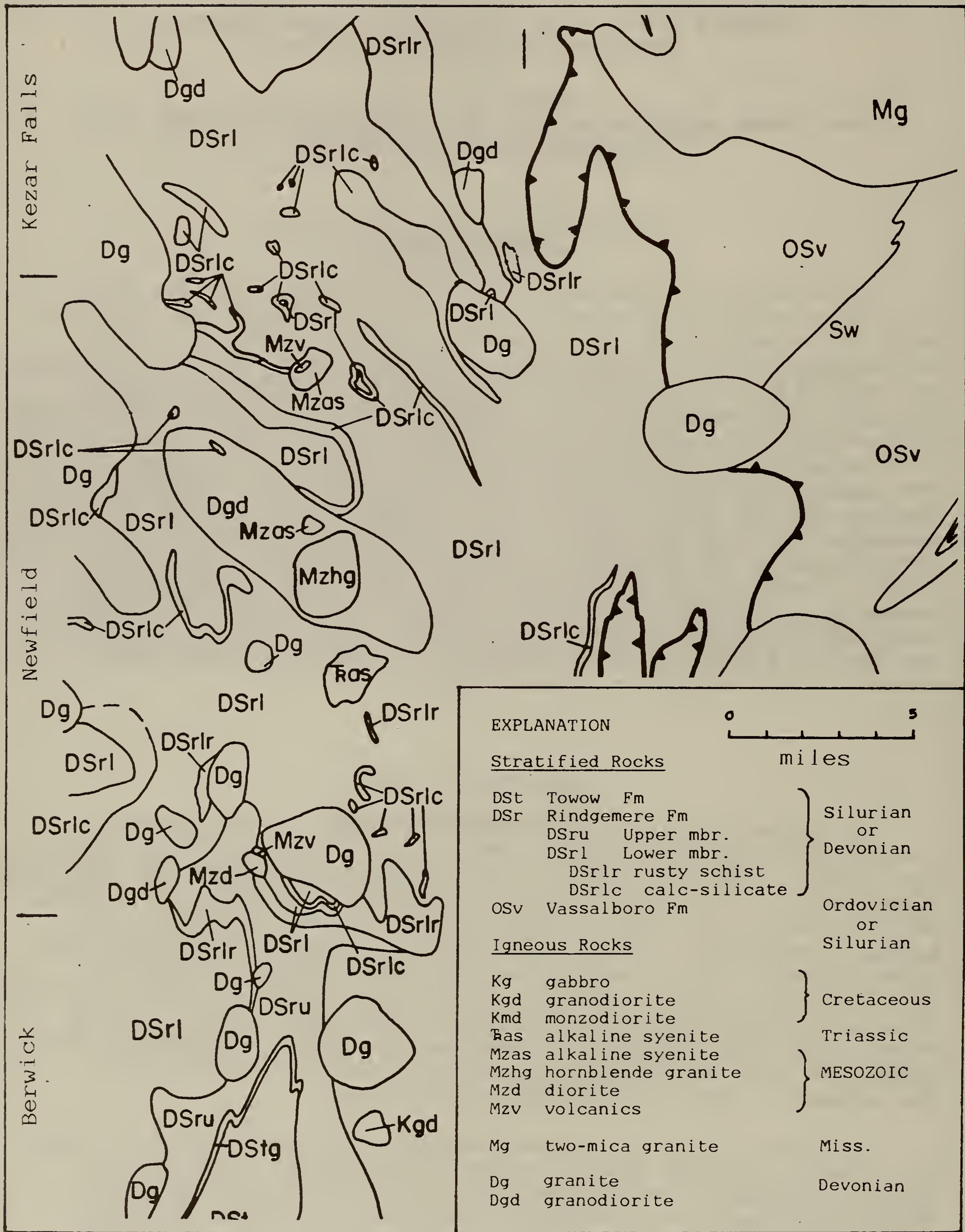

Figure 1. Bedrock map of parts of the Kezar Falls, Newfield and Berwick Quadrangles; from Hussey, 1985; mapping by Gilman (1977, 1978) and Hussey (1968).

monly found that steep southwesterly dipping beds are overturned. Evidence suggests a series of map-scale, doubly plunging northeasterly verging asymmetric folds with northwest-southeast trending hinge lines. This is in agreement with Eusden and others (1986) who suggest the existence of large scale recumbent folds that have subsequently been refolded into easterly verging asymmetric folds. Eusden and others (1986) further suggest that these northwest strikes in the Newfield area reflect yet an additional folding episode that deflects the regional northeast-southwest trends to the northwest-southeast trends observed. A structural analysis of fold geometries and fold generations for the area has yet to be completed.

The metasedimentary rocks have been intruded throughout the area (see figure 2) by plutons believed to be correlative with both the New Hampshire Plutonic Series and the White Mountain Plutonic-Volcanic Series (Billings, 1956). Rocks of the former consist of numerous stocks of light grey, binary granite lithically similar to the Sebago Pluton and others that have been dated at 322 ± 12 Ma (Haywood and others, 1984; Gaudette and others, 1982), and of bodies of fine-grained, medium grey, foliated granodiorite to quartz-diorite that resembles the Winnipesaukee quartz diorite. Both of these types are commonly cut by pegmatite. Five small stocks of White Mountain Plutonic Volcanic Series occur in the Newfield Quadrangle. These range in composition from quartz diorite (Acton Stock; seen on this trip, time permitting) to Na-rich syenites (Abbott Mountain Stock). They are non-foliated, and devoid of pegmatite that is so abundant in all other rock units. Two of these stocks have volcanic breccias associated with them (Gilman, 1983).

DESCRIPTION OF THE METASEDIMENTARY ROCKS

Non-bedded to poorly bedded mica schists and migmatites constitute the majority of the metasedimentary rocks in the region. In addition to the general red-brown or grey color and coarse-grained texture, the migmatites are characterized by abundant quartz-feldspar pods and lenses. The rocks may display a compositional layering ranging in thickness from a few millimeters to several centimeters. In most exposures large (5mm) porphyroblasts of muscovite can be seen oriented randomly across the schistosity. The foliation is typically highly contorted, and garnet and sillimanite are usually present. On the other hand, non- migmatized mica-garnet-sillimanite schists, in which muscovite and sillimanite lie in the plane of schistosity, tend to be less contorted and have a laminated appearance. Bedding (?), where present, shows up better on a weathered surface than on a fresh fracture, and typically ranges from a few millimeters to several centimeters. Both rock types appear to occur interlayered throughout the area and could not be mapped as separate units. They have been included

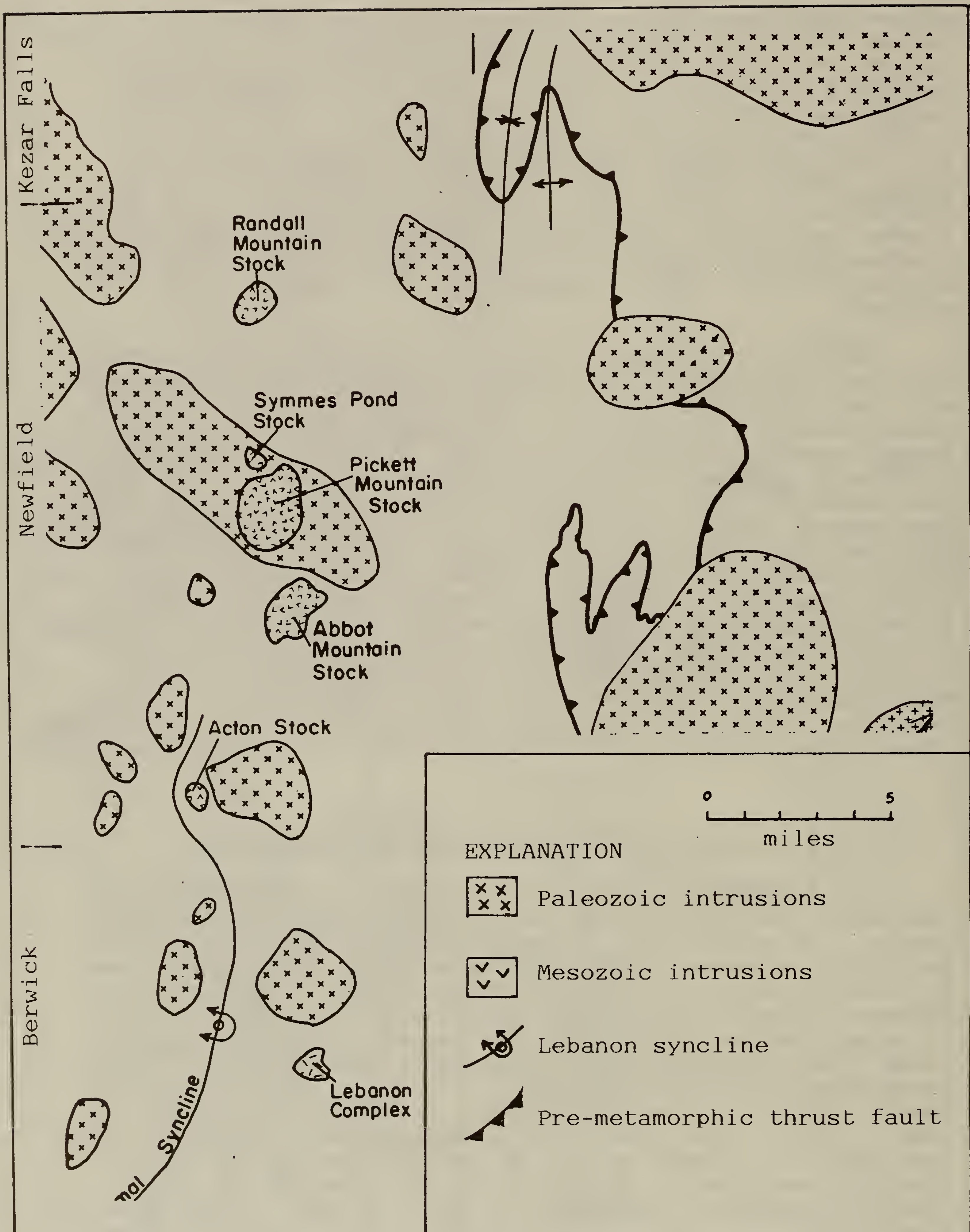

Figure 2. Intrusive rocks of the Kezar Falls, Newfield, and Berwick Quadrangles; from Hussey, 1985; mapping by Gilman (1977, 1978) and Hussey (1968).

within the Lower Rindgemere Formation by Hussey (1985) and Gilman (1977) and are perhaps the same as Eusden's unit 1 along the Spaulding Turnpike, which he correlates to parts of the Rangeley and Perry Mountain Formations of the Rangeley area. This type will be seen at stop 1 near the south end of Mousam Lake, and, time permitting, at the southeast end of Northeast Pond where it was mapped as unit 1 by Eusden (1984).

A well-bedded schistose quartzite is exposed in the Acton area and has been traced southward into the Berwick Quadrangle. Rhythmic bedding varies from 6 to 60 centimeters in thickness, the pelitic beds carrying conspicuous porphyroblasts of andalusite frequently pseudomorphed by muscovite. This was called Upper Rindgemere by Hussey (1985), and Eusden and others (1986) show it correlative in part to their unit 2 (Perry Mountain), and in part to unit 4 (Carrabassett). My mapping suggests that it can be traced through a series of isolated outcrops along strike into Eusden's unit 2. It will be seen at stops 6, 7 and 9.

Rusty weathering schists are found in two areas along the trip route. At stop 2, near the south end of Mousam Lake, the rock is poorly bedded, has an irregular schistosity, and weathers to a yellow-brown and black due to disseminated iron sulfide. On a fresh fracture the rock is light brown consisting mostly of quartz and muscovite. This occurrence appears to be a rusty layer within the migmatitic rocks although Eusden (1986) has included similar rocks within his unit 3. Other occurrences of rusty metasediments are known within the Rangeley Formation, and within Eusden's unit 1 north of Milton, New Hampshire. A second occurrence of rusty weathering schist will be seen at stop 8. This is a well foliated graphitic, crumbly, yellow weathering rock that Hussey (1985) calls Towow Formation, and which Eusden considers to be equivalent to his unit 3 (Smalls Falls).

Calc-silicate granofels is found in numerous isolated exposures in the Newfield Quadrangle and extends into New Hampshire in the vicinity of Great East Lake and Lovell Lake. The rock is characterized by its grey-green color, granular texture, and 1 to 10 cm. layering. It is fine-to medium-grained and contains thin interbeds of fine-grained, granular biotite-quartz-feldspar schist. Grossularite garnet and vesuvianite are occasionally abundant. Seldom are more than a few meters of calc-silicate exposed at any one outcrop, although the widths of the outcrop belts, particularly around Great East Lake would suggest a thickness of several tens to a few hundred meters. The exposure seen on this trip at stop 3 is one of the larger ones found in the area. I interpret it to be a layer within the migmatitic schists.

A summary of proposed stratigraphic correlations between Rangeley, Maine, southeastern New Hampshire, and this area is shown in table 1.

TABLE 1

Proposed stratigraphic correlations between the Rangeley area, southeastern New Hampshire, and the Newfield-Berwick Quadrangles.

AGE	Rangeley ME	Southeastern NH	Southwestern ME
	Seboomook	not exposed	not exposed
DEVONIAN	Hildreths	not exposed	Not exposed
	Carrabassett	Unit D4	Unit D4
	Madrid	Unit S4a	Unit S4a
SILURIAN	Smalls Falls	Unit S3	Towow
	Perry Mtn.	Unit S2	Upper Rindgemere
	Rangeley	Unit S1	Lower Rindgemere

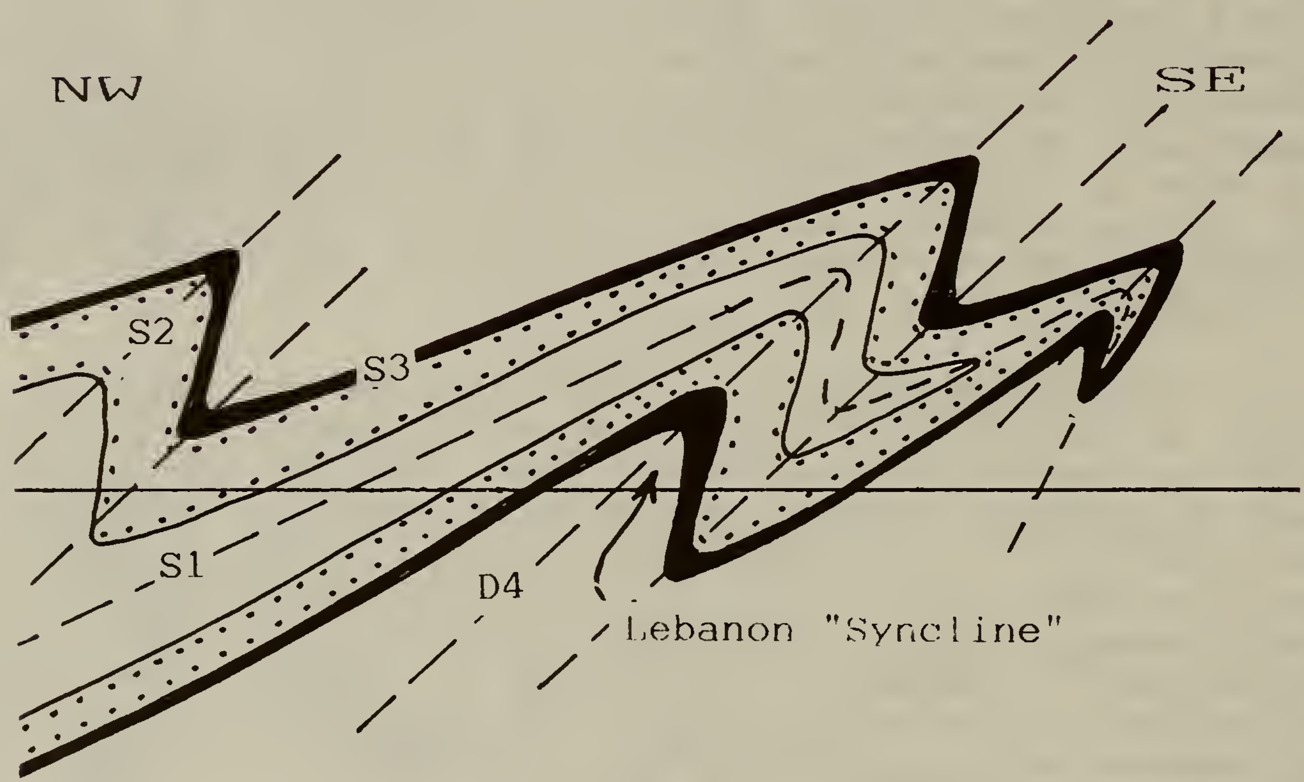

Figure 3. Schematic section of a refolded nappe; location of the Lebanon Syncline is indicated (after Eusden and others, 1986, figure 2-B). Units are the same as in figure 4.

STRUCTURE

Until Eusden (1984) studied the new exposures along the Spaulding Turnpike and worked out a stratigraphic section that turned out to be inverted on a regional scale, thus leading to his concept of a nappe structure in which this area constitutes the inverted limb (Eusden and others, 1986), mapping in the Newfield Quadrangle had not produced a regional structural pattern other than my conclusions that mesoscopic folds are folds of schistosity and therefore are at least second generation structures; that there is a tendency for these folds to be asymmetric with an easterly vergence, presumably in relation to the axis of the Merrimack Synclinorium to the west; and that folds in much of the Newfield Quadrangle trend northwest-southeast across the prevailing northeast regional structural trends.

Further development of a regional pattern was hindered by lack of exposure, the gentle dipping schistosity, and the lack of a known stratigraphic sequence. Hussey (1968) had mapped the Lebanon "Syncline", now reinterpreted as an antiformal structure in the inverted limb of a nappe (see figure 3), and I mapped the extension of this structure into the southern part of the Newfield Quadrangle. The nose of the structure lies just north of Acton. The structure is recognized by the presence of the non-migmatized, well-bedded schistose quartzite that Hussey (1985) calls Upper Rindgemere Formation, and by the rusty Towow Formation farther south in the Berwick Quadrangle.

Figure 4 is reproduced from part of Eusden and others (1986, figure 2) and is redrawn at the same scale as figure 5 from my mapping in 1985. While figure 4 provides a coherent pattern for the Lebanon "Syncline", it frequently conflicts with the distribution of rock units as I see them. Much of the difference stems from the interpretation placed on rusty units. I prefer to correlate the rock seen at Stop 6 with that found farther north at Acton even though some rustiness is encountered in between. Closing the fold at Acton as I show in figure 5 would then assign the rocks to the east (Stops 1,2, and 3) to unit 1 rather than unit 4 as shown by Eusden.

If the nappe model is correct, then we might expect to find the upper parts of the section exposed somewhere to the east where the nappe eventually closes. Well-bedded schists and quartzites are known to exist in the eastern part of the Newfield Quadrangle, but as yet their stratigraphic identity has not been determined.

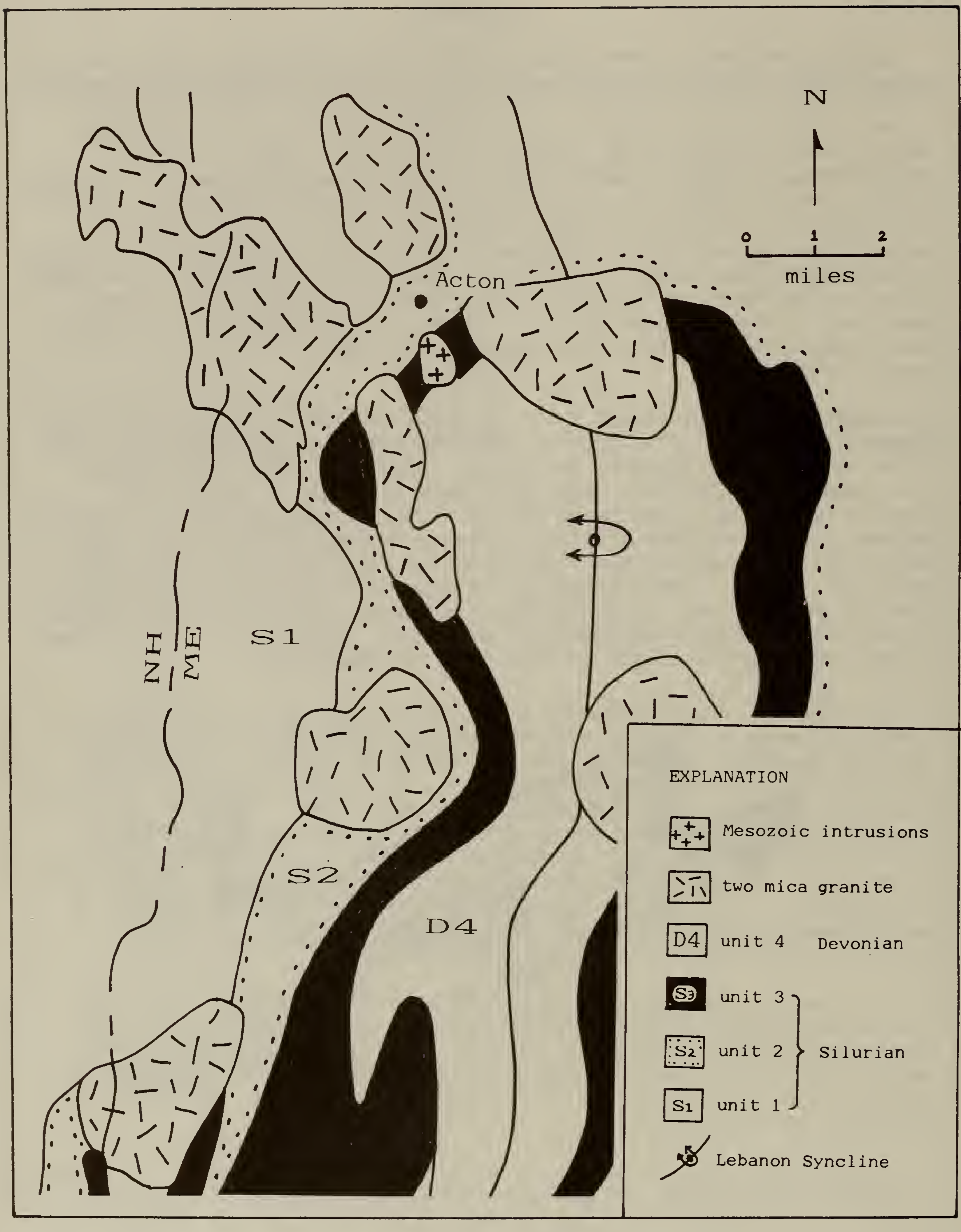

Figure 4. Geologic map of the north end of the Lebanon antiformal Syncline. (modified from Eusden and others, 1986, figure 2-A)

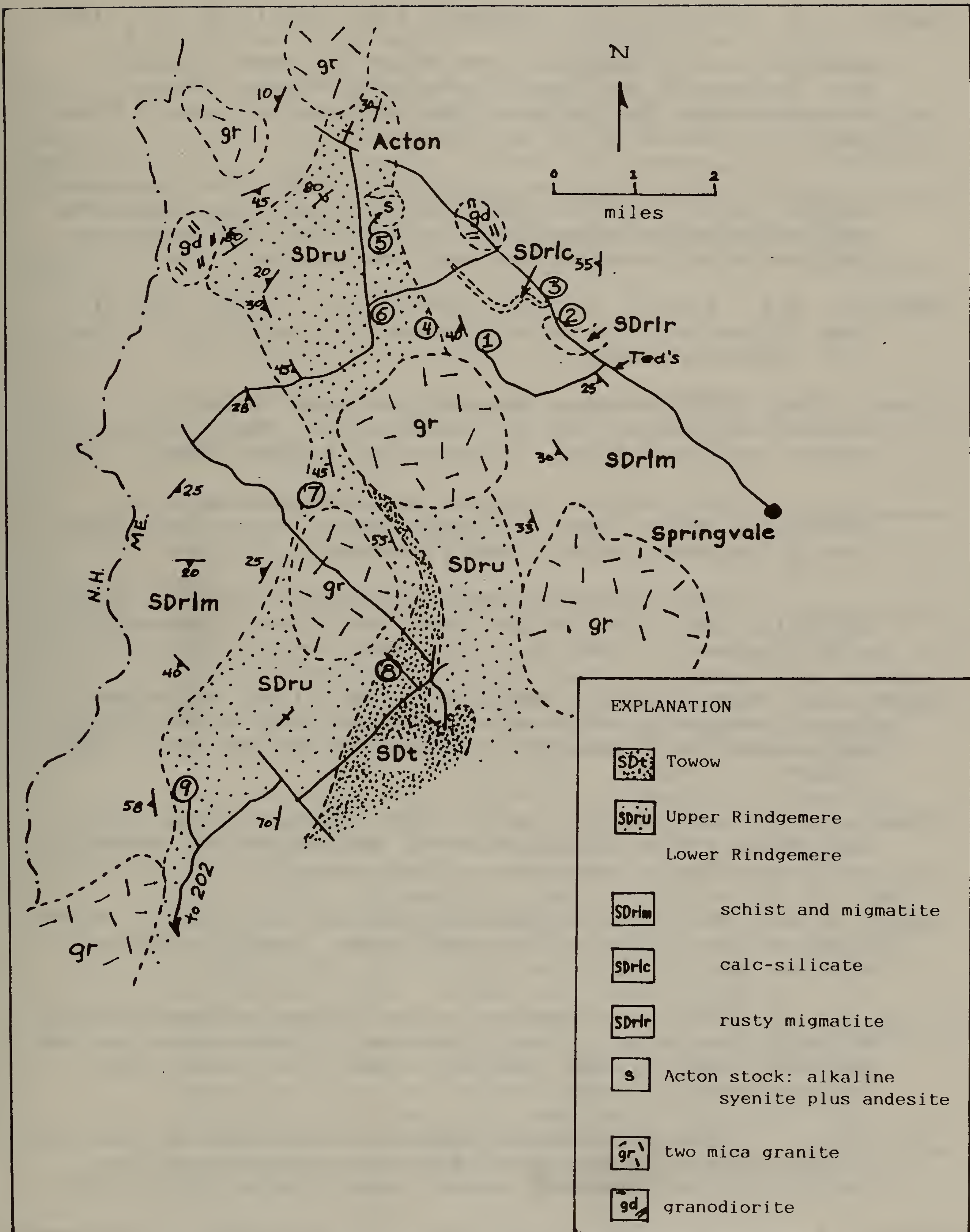

Figure 5. Geologic map of north end of the Lebanon antiformal Syncline (mapping by Gilman 1985-1986). Location of field trip stops are indicated.

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ITINERARY

Assembly point is Ted's Restaurant on Route 109 approximately 3 miles north of Springvale and 1 mile south of the Route 11 intersection near the south end of Mousam Lake. From Lewiston take the Maine Turnpike to Gray, Route 202 to Sanford, then Route 109 to Springvale. Route 224 northeast of Sanford will lead directly to Springvale and save a mile or two.

Meeting time: 9:30 a.m.: Allow 1 hour and 30 minutes driving time from the Lewiston toll gate of the Maine Turnpike to Ted's Restaurant.

Mileage	
0.0	Ted's Restaurant; proceed north on Route 109
0.1	Turn left.
0.5	Fork in road; bear right.
1.0	Fork in road; bear right (to 34th st. sign).
1.2	Fork in road; bear left on dirt road.
1.8	STOP 1: Pull off and park along main dirt road at end of Tattle Street (discontinued wood road that enters main road from left (west) at small angle; main road takes a sharp bend to the right up ahead); outcrop is 50 feet into the woods on the west side of road. Quartz-feldspar-biotite-muscovite-garnet migmatite. While some of these blocks are clearly not in place I think the top of the exposure is; at least the structural attitude agrees with other outcrops close by. Lenses and pods of quartz feldspar up to 3 cm. thick alternate with mica-rich layers of similar thickness. Small scale folds of the foliation are abundant but the overall attitude of the schistosity is N 25° W; 30°W. A few lenticular areas of fine-grained quartzite(?) are seen; these may be similar to the restite found within the migmatites along the Spaulding Turnpike (Eusden 1984). This is rather typical of the non-bedded schist-migmatite found throughout the Newfield Quadrangle and into the northern part of the Berwick Quadrangle. I think it is comparable to Eusden's unit 1 on the Spaulding Turnpike. Backtrack to Route 109.
3.5	Route 109; turn left.

4.2 Biotite schist and pegmatite on right.

4.3 Rusty on right.

4.4 <u>STOP</u> 2: Park on right shoulder.

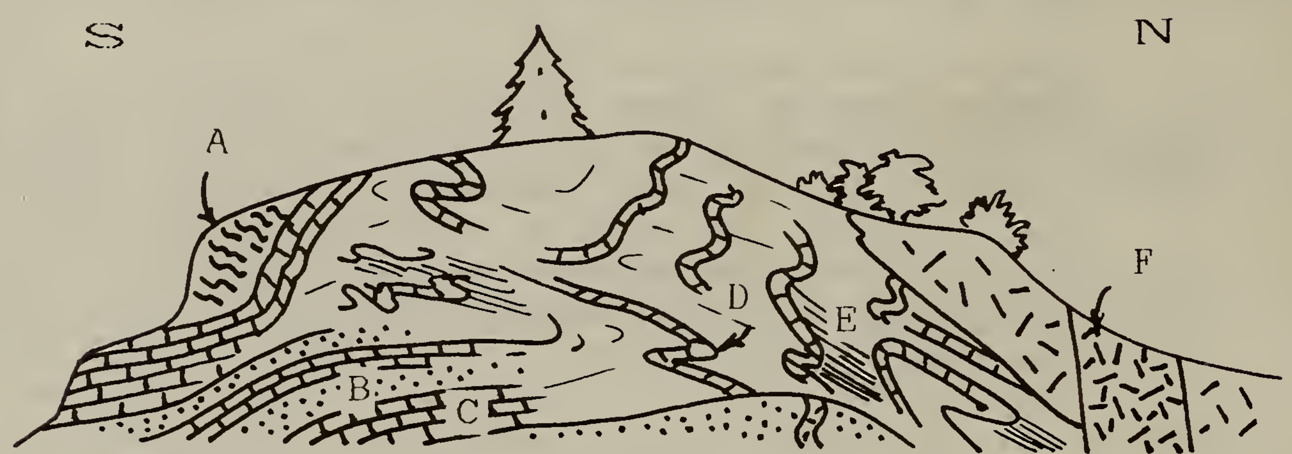
Rusty weathering quartz-feldspspar-garnet mica schist. Layering (bedding?) strikes N 20° W, dipping 25° W. Lenses of pegmatite are also present. Small isoclinal fold plunges S 30° W at 20°±. Scattered outcrops of this rock have been found to the east of here but there are no extensive exposures. Is this Smalls Falls; or is it a rusty unit within the migmatites such as that found along Route 16 north of Milton, NH?

Continue north on Route 109.

4.45 Junction with Route 11; bear left on Route 109.

4.5 Bridge and store; LAST PLACE TO BUY LUNCH.

4.7 <u>STOP</u> 3: Calc-silicate granofels with interbedded biotite schist and quartzite; also minor migmatite, pegmatite, and a diabase dike. Seen here is a complex disharmonic folding of grey-green calc-silicate with interbeds of biotite quartzite, which in some places is quite schistose. I have indicated particularly interesting features on the sketch below.



A. A small amount of migmatite is found at this end of the outcrop. Of special interest is the orientation of the layering which shows many small scale folds; in adjacent areas it is transposed into the axial plane direction of the small folds. The layering intersects the bedding of the calc-silicate at a high angle. This emphasizes the danger in assuming that foliation and bedding are always parallel.

B. Bed of massive biotite quartzite.

C. Thick bed of calc-silicate granulite with its characteristic grey-green layering. The general mineralogy of the rock is quartz-plagioclase-diopside, with occasionally abundant grossularite garnet and vesuvianite. The striking color banding is distinctive and easily spotted in outcrops. Exposures are usually rather small, this being one of the larger ones.

D. Excellent fold hinge in thin bedded calc-silicate and biotite schist. PLEASE; photograph, sketch, measure and otherwise enjoy it, <u>but don't</u> try to dig it out and add it to your collection! The geometry is that of a "Z-fold" with the hinge line trending S 45° W, plunging at 5-10°; its axial surface strikes N 45° E and dips 33° NW.

E. Well exposed fold hinges in the calc-silicate layer, and well developed axial plane cleavage in the adjacent biotite schist.

F. Diabase dikes are fairly common. I have not studied them in any detail, but in this area they tend to strike N 30° E, or east-west.

The granite and pegmatite are abundant throughout the area.

Continue north on Route 109.

5.5 Turn left on Milton Mills Road.

6.4 Bridge

6.45 Turn left onto old part of road and park. (Do not block road.)

<u>STOP</u> 4: 15 minute walk south along dirt road; approximately 1600 feet; take old wood road to the right. A 5 minute walk brings you to a clearing in a recently cut over area. The outcrop is high up to your right (a 25 foot cliff).

Well bedded biotite schist and quartzite; layer contacts are sharp and non-graded(?). This rock structurally overlies the migmatitic rocks down slope to the east, much like those seen at Stop 1. Bedding thickness ranges from 0.5 cm. to 15 cm.; quartzite and pelite both of about equal thickness. The top of the slope on the southern end of the exposure presents an excellent cross-sectional view of a series of asymmetric folds with an easterly vergence. Axial surfaces strike N-S and dip 55° W.

I consider this to be transitional between the migmatites to the east and the well-bedded Upper Rindgemere found to the west at Stop 6.

Return to cars and continue west on Milton Mills Road.

7.0 Baptist Church on left; fork in road:

For optional Stop 5, bear right and follow the road log below. To omit Stop 5, bear left and continue 0.3 miles to Stop 6.

For optional Stop 5;
0.1 intersection, turn right.
1.3 turn right through gate into apple orchard.

<u>STOP</u> 5: Acton Stock. In the parking area there are several pavement exposures of well-bedded quartzite and schist; this is the Upper Rindgemere Formation that I correlate with the rock seen at Stop 6. The contact zone with the stock is seen in small pavement outcrops near a pile of pipe; here a fine phase of the quartz diorite encloses breccia fragments of hornfelsed metasediment. The interior part of the stock is found in the wooded area about 100 yards to the southeast. The rock is medium grained, medium-to-dark brown on a fresh surface, and has a vertical jointing striking N 70° E. The rock consists of approximately 80% plagioclase (An_{50}), 15% pyroxene (both ortho and clino), and 5% quartz, apatite and opaques.

A dark grey, porphyritic and fragmental andesite is exposed in the woods to the north of the orchard.

From the orchard gate turn left, return south to Stop 6.

2.3 road on left; go straigt ahead.
2.4 Stop sign.; bear right.
2.5 Stop 6.

7.4 <u>STOP</u> 6: Exposure of Upper Rindgemere in the stream on the east side of the road.

Well-bedded quartzite and schist with conspicous andalusite porphyroblasts; strike N 15° W, dip 30° E. A nearly horizontal crenulation lineation trends N 20° w, and is related to a spaced cleavage that dips 40°W. Bedding thickness ranges from laminations to 50° cm.; contacts between schist and quartzite are sharp with no definitive tops.

This unit can be traced northward as far as Acton where I think it ends in the nose of a fold. The rock appears to lie structurally above that at Stop 4.

For those of you familiar with the Rangeley section, is this Perry Mountain or Carrabassett, or neither?

8.3 Road to left and Lincoln Schoolhouse; continue straight ahead.

8.7 Bear left at fork in the road.

8.9 Intersection; continue straight ahead.

10.4 Miller Corner intersection; turn left on Hilton Ridge Road.

12.0 Turn left onto gravel driveway.

STOP 7: Contact (almost!) between migmatites cropping out in the dooryard and well-bedded quartzite and schist in the pasture. Migmatite is extremely contorted and occasionally shows disrupted beds of biotite quartzite. "Lumps" may have been andalusite but are now muscovite.

In the pasture well-bedded quartzite and schist strike north-south, dipping 80° W. A subtle foliation strikes about N 25° E. Facing direction from graded bedding is unclear. However the outcrop near the rusty hay rake has a fold hinge, so tops directions may be meaningless on a large scale. The bedding style (quartzite 6 to 25 cm., schist up to 50 cm.)., and the abundance of andalusite pseudomorphs are similar to Eusden's unit 2 along the Spaulding Turnpike. Hills to the west are all underlain by migmatite similar to that seen at Stop 1.

12.2 Back at entrance to driveway; turn left.

12.3 Migmatite in yard on right.

13.2 Two-mica granite (Sebago type).

15.2 North Lebanon; bear right on Heath Road.

15.3 Woodman's Used Cars on left.

15.4 Turn right.

15.8 Turn left into driveway.

STOP 8: Towow Formation exposed in small pond. Bedding and schistosity strike N 20° E, dipping 50° W.; 6 cm.± beds of quartzite alternate with layers of black graphitic, crenulated phyllite. North along this road there are additional exposures of rusty weathering pelite with andalusite pseudomorphs, then

non-rusty quartzite and pelite.

16.2 Backtrack to Heath Road and turn right.

18.2 Center Lebanon; turn right on Center Road.

18.5 Turn left.

19.3 Road intersection from left; go straight ahead.

19.9 Turn right on dirt road.

20.3 Turn left into Gully Oven parking area.

STOP 9: Gully Oven; a spectacular exposure of well-bedded quartzite and schist. This is Eusden's unit 2 (Perry Mountain?), and I suggest it is correlative with the rock seen at stops 6 and 7. There are several excellent structural features displayed here, bring your camera and sketchbook.

Bedding strikes N 10° W, dipping 45° W; the beds show fast grades, that is, contacts on both sides of a sandy bed are quite sharp, but on close inspection you should be able to find some that indicate the beds are overturned. Sandy beds range in thickness from one to several centimeters; shaley layers are much thicker, particularly on the west side of the valley.

In the narrowest part of the "canyon" quartz+garnet cotecules occur as pods and thin layers. These have a pinkish color and a fine-grained, granular texture.

In the same place a thick pegmatite layer has been boudinaged in the plane of the schistosity. Also a thin pegmatite vein displays "texbook" ptygmatic folds with "S" and "Z" folds as well as boudinage.

Just upstream and on the opposite side, in the side of a pothole, several fold hinges of an isoclinally folded sandy layer are visible. Notice that the schistosity is axial planar to these folds. Also notice that schistosity bows around the andalusite porphyroblasts, indicating that it has been flattened following the growth of the andalusite.

The schistosity and bedding in fact are not quite parallel; they are about 5° off, the bedding being the steeper. In the side of another pothole up near the parking area, there is an excellent example of graded bedding crossed by axial plane schistosity (cleavage).

END OF TRIP: Backtrack to tar road; turn right and proceed through West Lebanon to Route U.S. 202 (about 5 miles.)

TRIP C-4

CARBONIFEROUS METAMORPHISM ON THE NORTH (UPPER) SIDE OF THE SEBAGO BATHOLITH

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INTRODUCTION

Until recently the high-grade metamorphism in western Maine was routinely accepted as Devonian in age. Initially, only a single Devonian event was assumed and Pankiwskyj (1965) appears to have been the first worker to tentatively suggest the existence of more than one Devonian metamorphism. Subsequently, Guidotti (1970) described explicitly at least three Devonian metamorphisms. The most recent overviews of the Devonian metamorphic events in central and western Maine have been given by Holdaway et al. (1982) and Guidotti et al. (1983). Through this time there were only vague suggestions of possibly younger high-grade regional metamorphic rocks, Osberg (1968) and Guidotti (1970).

The first really suggestive evidence for a post-Devonian high-grade metamorphism in western Maine was the independent radioactive age determinations of the Sebago batholith (see Fig. I) by Hayward and Gaudette (1984), Aleinikoff (1984), and Aleinikoff et al. (1985). These studies indicated that the S-type granite of the Sebago Batholith has a crystallization age of 325 Ma. Moreover, Aleinikoff et al. (1985) briefly discussed the possibility that the high-grade metamorphism surrounding the pluton might have the same age. Inasmuch as the abundant metapelites around the Sebago pluton are migmatitic and much intruded by pegmatites that seem to be derived from the Sebago body, this possibility seemed to merit further consideration.

The convincing evidence for this suggestion was provided by Lux and Guidotti (1985). They showed that:

(a) Although the Songo pluton (Fig. I) has a crystallization age of 382 Ma (Lux and Aleinikoff, 1985), the hornblendes in it were re-set so that they passed below their blocking temperature (Ca. 500^{0}C) at about 308 Ma.

(b) Hornblende in the southern portions of the Mooselookmeguntic pluton (Fig.1) have strongly disturbed Ar spectra but hornblende from the northern part gives ages close to the pluton's crystallization age (371 Ma, Moench and Zartman, 1976).

(c) The areal distribution of the K-feldspar + sillimanite isograd as mapped in the Bryant Pond quadrangle (Evans and Guidotti, 1966) and in the Buckfield quadrangle (Guidotti et al. 1973) is spatially related to the outcrop pattern of the Sebago batholith.

(d) Textural features were noted (see below) which strongly suggested

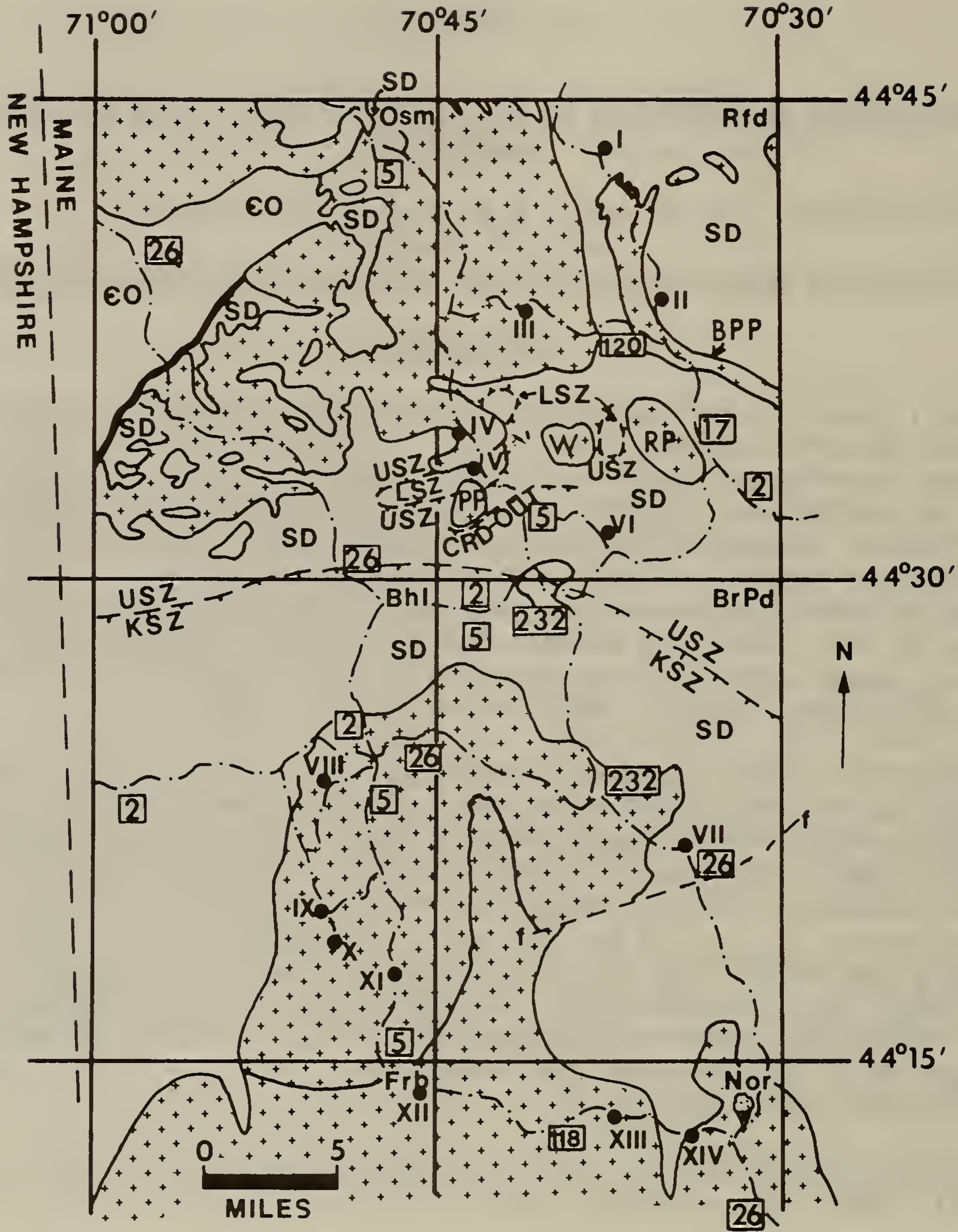


Figure 1. Roadlog and location of plutons and metasediments. €O - Cambro-Ordovician strata; SD - Siluro-Devonian strata. Dot-dash lines - roads; Boxed numbers - State routes. Roman numerals - localities visited. Hachured line - isograds; LSZ, USZ, KSZ - lower sillimanite, upper sillimanite, and K-feldspar + sillimanite zones; CRD-out - breakdown of cordierite. Heavy black line - boundary of €O and SD strata. f-fault. PP, W, RP, and BPP - Plumbago, Whitecap, Rumford, and Bunker Pond Plutons. (+) Granitic bodies differentiated on Figures (2) and (3). Osm, Rfd, Bhl, BrPd, Frb, Nor - Old Speck Mountain, Rumford, Bethel, Bryant Pond, Fryeburg, and Norway 15' quadrangles.

post-Acadian recrystallation in the central and northern portions of the Rumford quadrangle.

Subsequently, thermal modelling (DeYoreo et al. 1985) in the context of the Sebago pluton being a thin (ca. I km thick; Hodge et al. 1982) north-dipping sheet strongly supported the suggestion of the pluton being the heat source for the areally extensive Carboniferous metamorphism suggested by Lux and Guidotti (1985).

GOALS OF THE FIELD TRIP

The goals of this field trip are aimed at the following:

(a) To inspect the igneous and metamorphic rocks that are involved in the Carboniferous metamorphism

(b) To discuss, in the context of the field outcrops, the nature of the Carboniferous metamorphism and the evidence for it.

(c) To discuss some interesting new mineralogic data for the Songo pluton which may be related to the metamorphism caused by the Sebago batholith.

(d) To elaborate on some details of a model for the thermal regime established by intrusion of the Sebago batholith.

GENERAL GEOLOGIC SETTING

The area involved in this discussion of Carboniferous metamorphism includes the Rumford, Old Speck Mountain, Bryant Pond, Bethel, and the northern parts of the Norway and Fryburg 15' quadrangles, see Fig. I. Geologic control has been provided by the mapping of Milton (1961), Moench and Hildreth (1976), Guidotti (1965) and Fisher (1962) and the new State map (Osberg et al. 1985).

All of the meta-sediments of concern are Siluro-Devonian age and lie within the Merrimack Synclinorium. Most are meta-pelites but some biotite granulites, conglomerates and calc-silicates are also present. The strata are tightly folded, mainly into NE trending folds, but some NW trending folds are also present in the Bryant Pond quadrangle.

A great variety of igneous rocks is present, most of which are granitoids and range from tonalite to granites, though some small gabbroic bodies are observed. The major granitic plutons are shown on Fig. I and those discussed specifically herein are named. Also present are very abundant pegmatites and aplites, especially in the Bryant Pond and Bethel quadrangles and southward. Generally they increase in abundance as the Sebago batholith is approached and in many areas the metamorphic country rocks are merely screens and inclusions amounting to about 50% of the total rock. (e.g. see Guidotti, 1965).

INTRUSIVE ROCKS

A major part of the area affected by the Carboniferous metamorphism associated with the Sebago pluton is composed of two large intrusions - the Mooselookmeguntic and Songo plutons. Indeed much of the evidence for this metamorphic event comes directly from the igneous rocks. Within these plutons the rock types vary from quartz - diorites and tonalites to two mica leucocratic granites and span the meta - to peraluminous compositional range. The field and petrographic characteristics of these plutons (Mooselookmeguntic, Songo and the Sebago) are detailed below along with geochemical data for the Songo. In addition, two other important aspects - the geometry of the plutons and their geochronology - are also discussed as they have particular relevance to the Carboniferous metamorphism observed in Western Maine.

Field Relations and Petrography.

The Mooselookmeguntic pluton exhibits a wide variety of igneous rock types and thus is probably a composite intrusion. Various phases have been mapped by Moench and Hildreth (1976) in the Rumford quadrangle and these are also observed in adjacent areas to the N and W (i.e. in the Old Speck Mt., Oquossoc and Rangeley quadrangles), Fig. 2.

The major rock type in the northern part of the pluton (Rangeley and Oquossoc quadrangles) is a light colored, two-mica granite which has a medium grained, equigranular texture. Some small garnet euhedra are observed in this rock indicating its peraluminous character, and Apatite and zircon are the main accessory phases. This two mica granite is also observed in southern parts of the pluton as seen at Locality 4 of the fieldtrip.

A granodiorite - tonalite rock type forms the major part of the Mooselookmeguntic pluton along part of its eastern contact (Fig. 2). This is usually a medium grained granodiorite which is often foliated and contains abundant hornblende (in addition to biotite) and sphene. Accessory phases present are allanite, zircon and epidote. In some parts of this granodiorite metasedimentary xenoliths are abundant as is observed close to the outer contact of the Mooselookmeguntic in exposures east of Little Ellis pond.

Much of the central zone of this pluton is mapped by Moench and Hildreth (1976) as a two mica granite which contains abundant inclusions of hornblende and sphene-bearing granodiorite (Fig. 2). Such inclusions are often several meters across and their foliation is often truncated by the presumably later two mica granite. Both of these rock types are observed at Ellis Falls (Locality 3 of fieldtrip) and there is a clear distinction between the dark gray mafic rich granodiorite blocks and the fine grained, leucocratic two mica granite.

The field and petrogenetic relationships between the various phases of the Mooselookmeguntic pluton are at present unclear but it is hoped that ongoing research will clarify the composite nature of this intrusion.

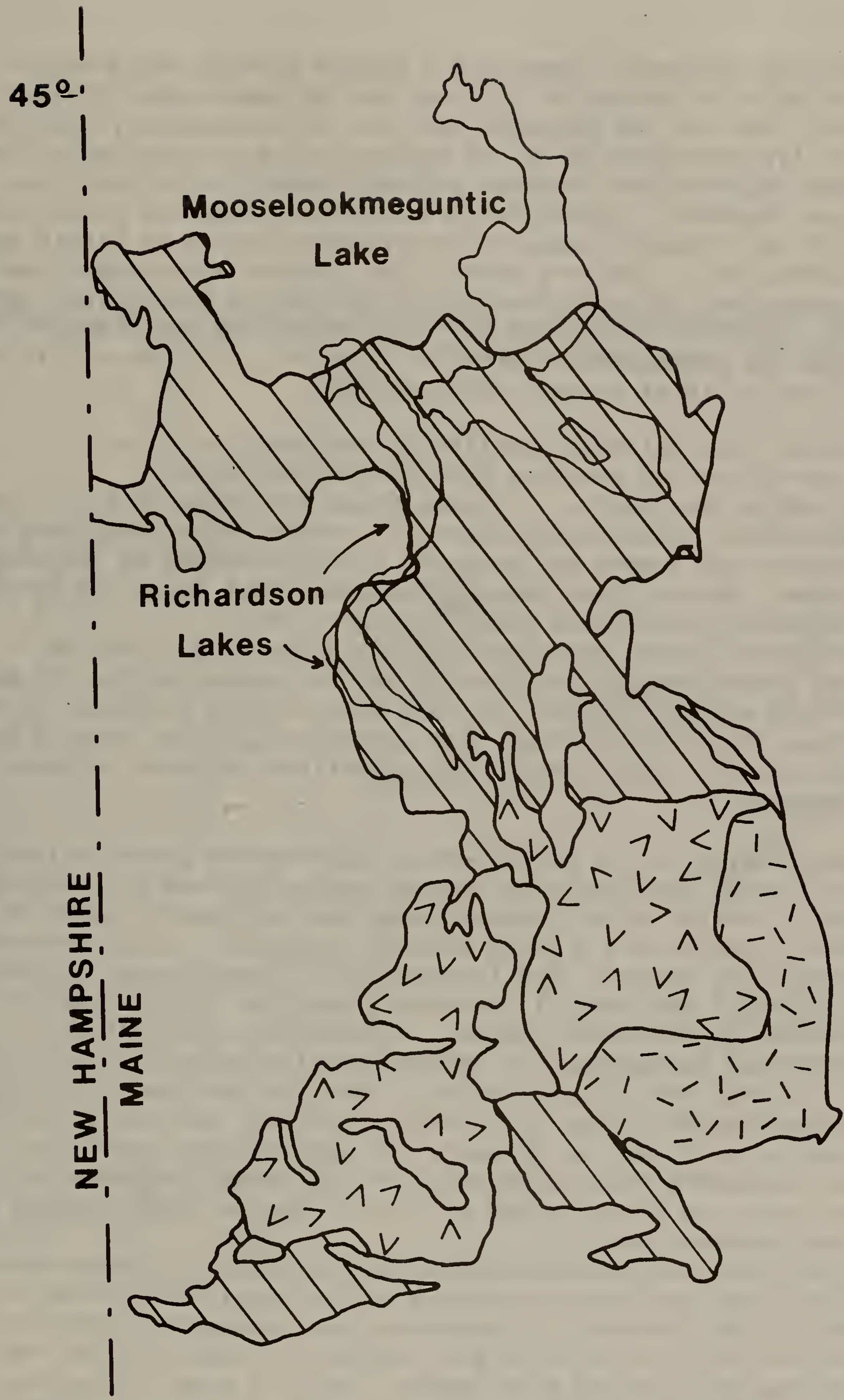


Figure 2. Map illustrating the various phases of the Mooselookmeguntic pluton; diagonal lines = two mica granite, / = hornblende, sphene granodiorite and V = two mica granite with abundant inclusions of granodiorite; unshaded areas within the pluton are large country rock inclusions. After Moench and Hildreth (1976).

The Songo pluton is generally a medium grained equigranular granodiorite which in places is foliated and in some cases intensely banded. Xenoliths are not abundant but some metasedimentary enclaves of local origin (?) are observed at its margins and also internally (Locality 10). Igneous enclaves are uncommon although some biotite knots and schlieren are observed. Pegmatites frequently intrude the Songo and are observed both as irregular masses atop prominent knolls or bluffs and as dike-like intrusions. They are usually of a simple mineralogy (quartz, feldspar, muscovite) although some do contain beryl, tourmaline, garnet and apatite. Commonly associated with the pegmatites are areas of fine grained two mica, garnetiferous granite (Locality 11). The origin of these small bodies is at present problematical.

Initial reconnaisance sampling of the Songo for the $^{40}Ar/^{39}Ar$ dating studies revealed distinct mineralogical variations within this pluton. In some areas the Songo is a biotite + hornblende + sphene granodiorite whereas in other parts the granodiorite lacks hornblende and sphene and contains Ti-rich biotite as the only mafic silicate phase. Recent field mapping has shown that these two Songo varieties crop-out in discrete areas (Fig.3) - the biotite-hornblende-sphene granodiorites being observed in the NW, N and E parts of the pluton whereas the Ti-rich biotite granodiorites are exposed in the southern areas and appear to loop around the N extension of the Sebago pluton. It is significant that within this latter area of the Songo, foliations are more intense and pegmatites are more abundant and areally extensive.

Mineralogically the biotite-hornblende-sphene granodiorites are obviously an I-type (metaluminous) assemblage as defined by Chappell and White (1974). Biotite always predominates over hornblende which is absent in some areas, plagioclase greatly exceeds K-feldspar in abundance and the latter is typically anhedral and interstitial in appearance. Sphene is relatively abundant and commonly euhedral. Apatite, zircon and allanite are also present as accessory phases. In contrast the biotite only granodiorites lack hornblende and sphene, apatite being the dominant accessory (+ zircon and minor allanite), the biotites have a distinctive red-brown pleochroism typical of Ti-rich varieties, and muscovite is present, some of which may be primary. In thin section quartz and plagioclase frequently show evidence of being strongly deformed e.g. plagiocalse twins are often warped and quartz grains show undulatory and sectorial extinction.

Usually the two Songo varieties can be easily identified in the field either by the presence of hornblende and sphene or a red-brown sheen of the biotites (e.g. at Locality 9). However in some parts of the Songo, Ti-rich biotites do co-exist with sphene. In this case the sphenes are usually anhedral and the biotites do not show the extreme red-brown pleochroism evident in rocks in which sphene is totally absent. Therefore, there may be a gradation between the two Songo varieties. Hence, no definite contact is shown on Fig. 3.

Modal analyses reveal that most samples from both Songo varieties plot in the granodiorite field of Streckeisen's classification diagram (Fig. 4) although some are strictly speaking, tonalites. Importantly,

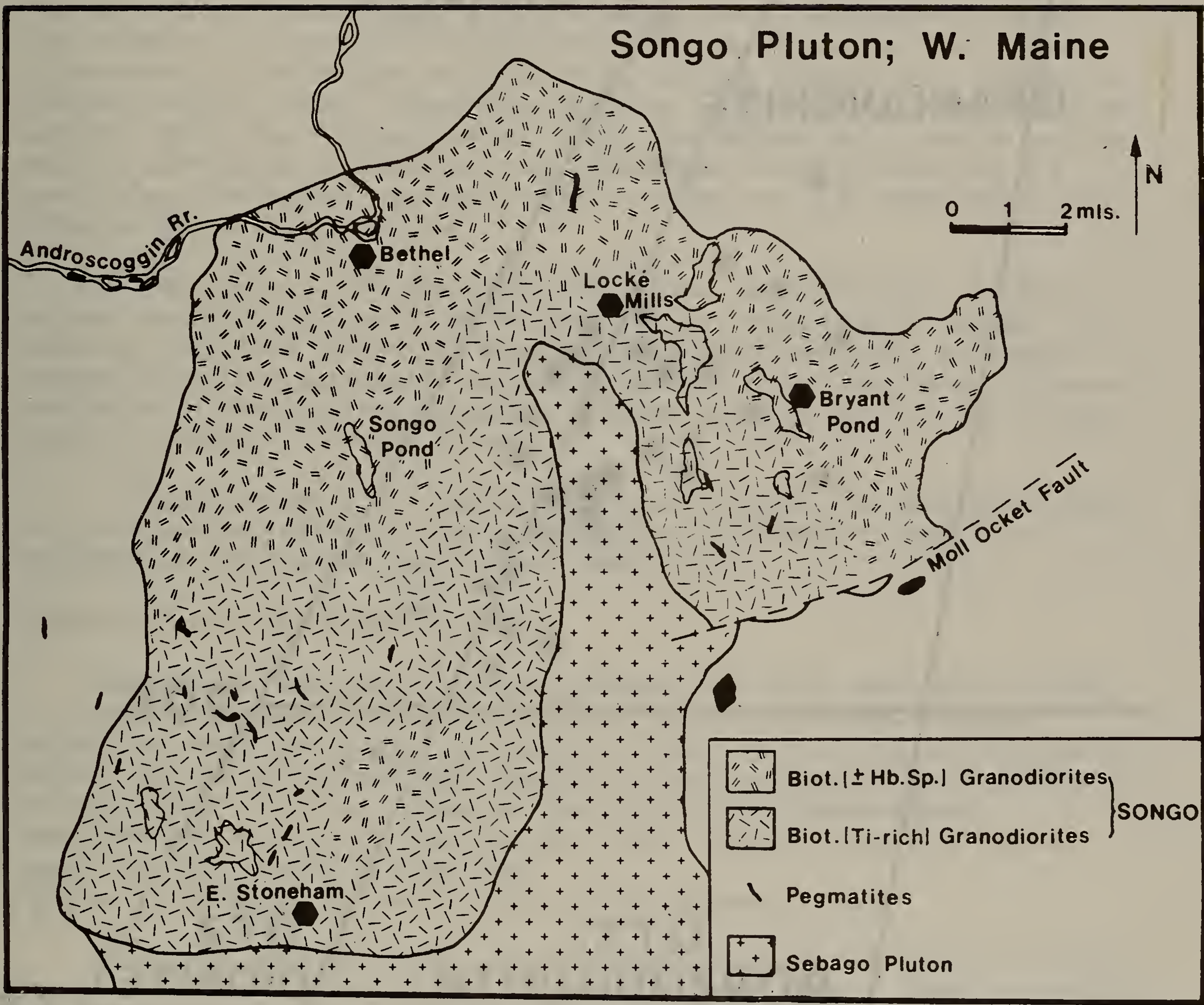


Figure 3. Generalized map of the Songo pluton showing the disposition of the biotite + hornblende + sphene granodiorites and the Ti-biotite granodiorites.
Also shown are some of the larger pegmatite bodies as well as the N part of the Sebago pluton.

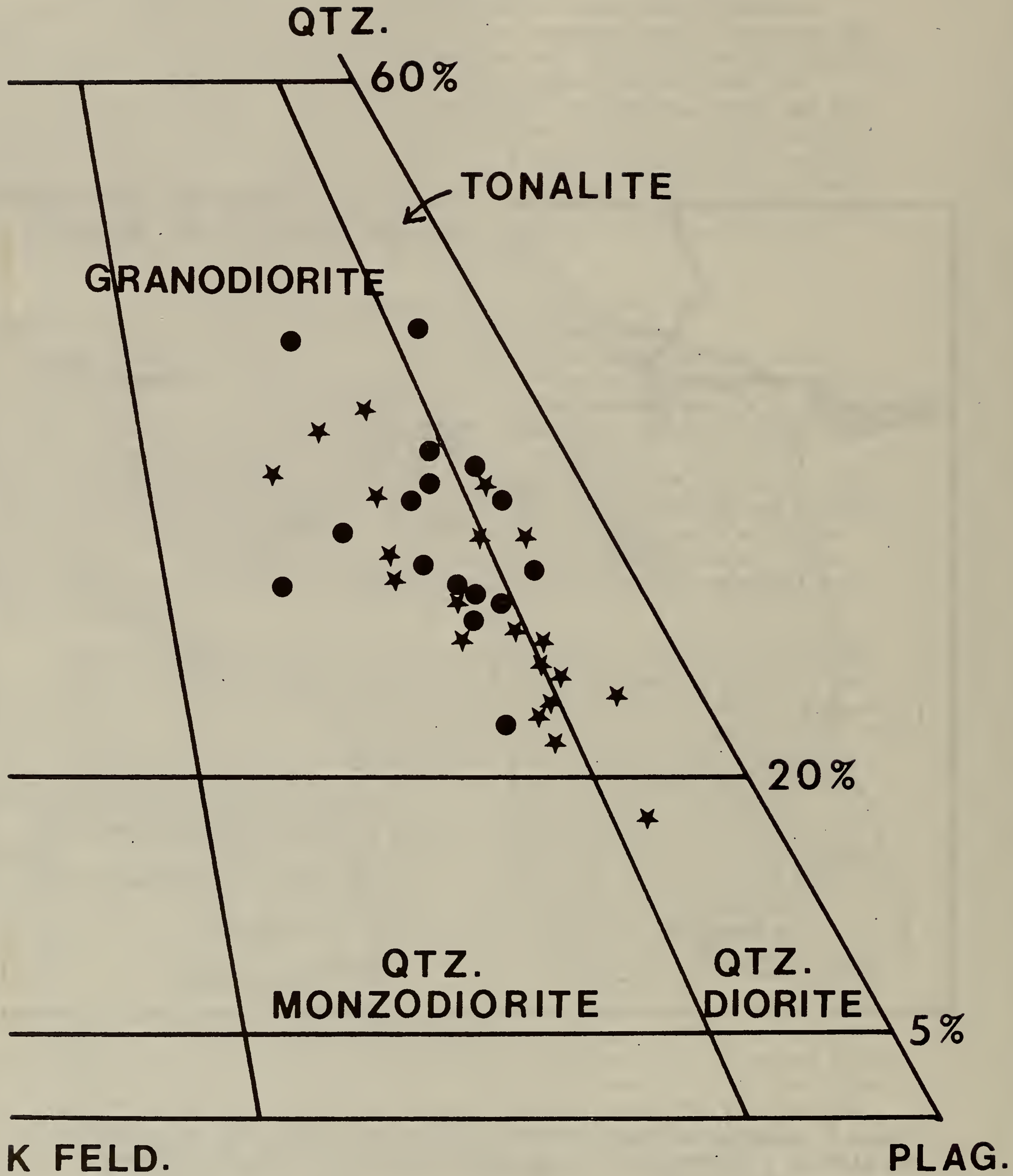


Figure 4. Streckeisen's Q-A-P classification diagram showing modal data for the biotite - hornblende - sphene granodiorites (stars) and the Ti-rich biotite granodiorites (filled circles) of the Songo pluton.

neither variety can be distinguished using these modal parameters and there is a large degree of overlap between them.

In summary, the Songo exhibits distinct internal mineralogical variations -- parts of the pluton are a typical I-type granodiorite but significant areas do appear to have a peraluminous 'S-Type' affinity.

The granite of the Sebago pluton is essentially a medium to coarse grained equigranular rock, although finer grained examples are observed. Two varieties of the Sebago, pink and white, were recognized by Hatch (oral commun. to Aleinikoff et al., 1985) but these may just be due to differences in the amount of feldspar alteration (Berry, oral commun. to Aleinikoff et al. 1985) rather than separate phases of the pluton. Recent mapping (State Map - Osberg et al. 1985) has recognized a zone along the eastern contact of the Sebago in which metasedimentary xenoliths are particularly abundant. Similar, smaller areas are also observed in the internal parts of the pluton.

The Sebago is of definite peraluminous affinity with abundant primary muscovite, in addition to biotite, and occasionally garnet. The ratio of muscovite to biotite does change across the pluton although whether this is a systematic variation has yet to be determined. Commonly the color of the biotites suggest a Ti-rich variety. Apatite and zircon are the dominant accessory phases.

Numerous basic to intermediate dikes intrude the Sebago and many of these may be related to the younger alkaline Mesozoic stocks as observed on Rattlesnake and Pleasant mountains.

Geochemistry of the Songo pluton

Geochemical analyses of Songo granodiorites were performed (by XRF) in order to elucidate the origin of this pluton given the internal variations described above. A summary of some of the results is given in Table 1 and it is obvious that there is a large degree of overlap between the two Songo varieties. However the biotite-hornblende-sphene granodiorites do tend to have the more mafic compositions with higher levels of TiO_2, Fe_2O_3(T), CaO and P_2O_5 wt. % compared to the Ti-rich biotite granodiorites.

Bivariate diagrams (e.g. Fig. 5) also illustrate the extent to which the two Songo varieties overlap in chemical composition. Furthermore all the samples define single, essentially continuous compositional trends with TiO_2, Fe_2O_3(T), MgO, CaO and P_2O_5 wt.% all decreasing as SiO_2 wt% increases. However there is a greater degree of scatter on some plots among the Ti-rich biotite granodiorites. It may also be significant that Na_2O and K_2O wt. % as well as some trace elements (e.g. Ba ppm) do not show any definite trends with increasing SiO_2 content.

In discussing the origin of the Songo pluton it is important to take into account the distinct mineralogical variations evident in this intrusion and the close spatial relationship between the Ti-rich biotite granodiorites and the later Sebago granite. Furthermore, the geochemical

TABLE 1.

Summary of Major/Minor Element Results (wt. %) for the Songo Pluton in the form of mean (top line) and range values (in parentheses). Analyses are by XRF.

	Biot. (± Hornblende, Sphene) Granodiorites (N = 19)	Biot. (ti-rich) Granodiorites (N = 18)
SiO_2	63.59 ± 2.5 (57.28 - 67.72)	66.01 ± 2.2 (63.36 - 70.9)
TiO_2	0.804 ± 0.12 (0.643 - 1.027)	0.698 ± 0.13 (0.404 - 0.926)
Fe_2O_3(T)	4.45 ± 0.71 (3.28 - 5.87)	3.92 ± 0.65 (2.48 - 5.06)
CaO	4.18 ± 0.50 (3.36 - 4.91)	3.78 ± 0.56 (2.69 - 4.55)
P_2O_5	0.296 ± .10 (0.179 - 0.652)	0.241 ± 0.058 (0.129 - 0.329)

variations described above are also relevant in discussing the petrogenesis of the Songo granodiorite. Possible mechanisms which should be considered are i) the mixing of an I-type magma and an anatectic melt, ii) the fractional crystallization of an I-type magma which has been progressively contaminated with a sedimentary component and iii) a process of restite unmixing (White and Chappell, 1977). However, field and petrograpic evidence for the latter process is largely absent from the Songo. Further critical assessment of these petrogenetic mechanisms must await future Sr and O isotopic analyses.

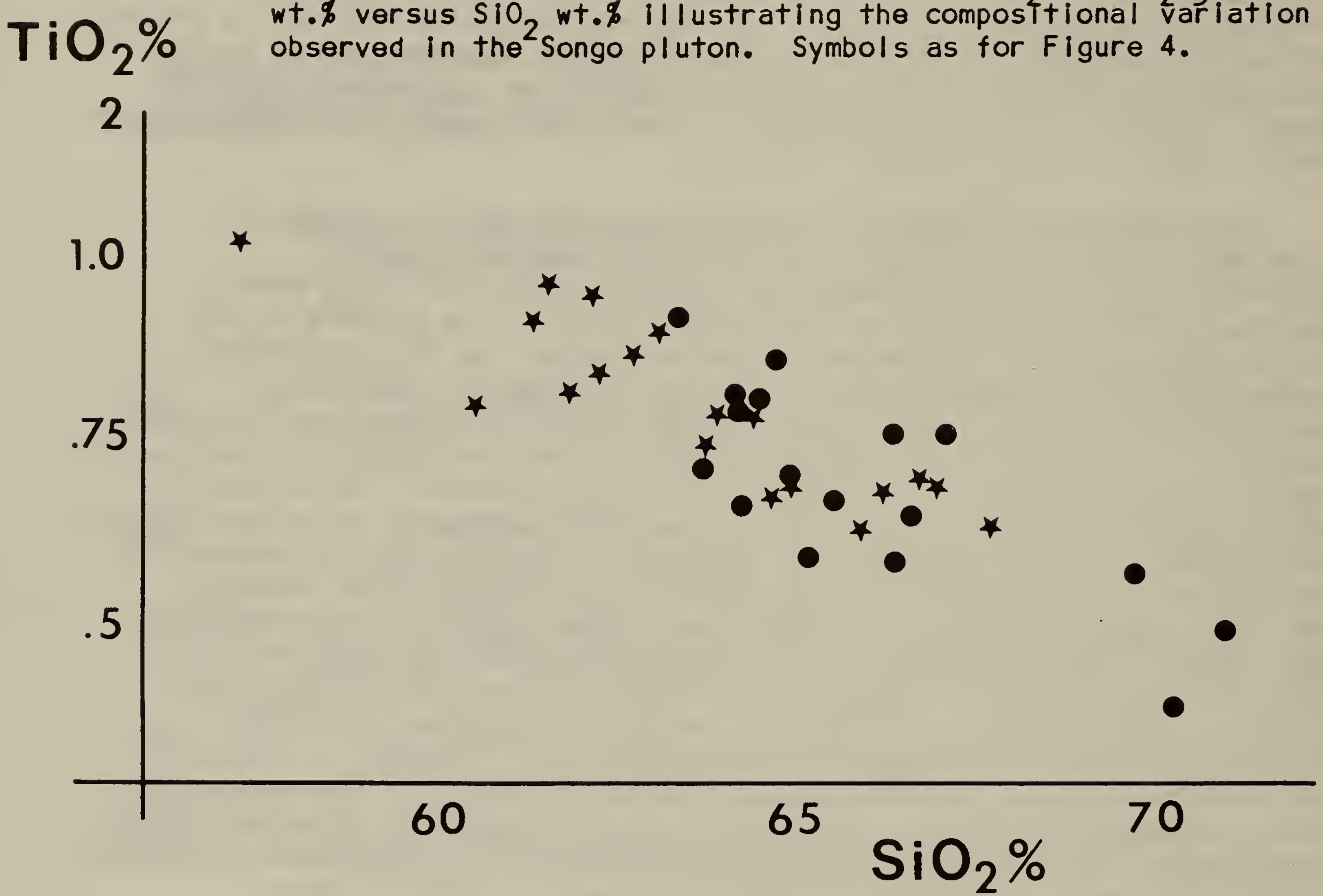
Apart from these 'magmatic' processes it is also important to consider the possibility that the Songo has been modified by post-crystallization processes (i.e. reheating and fluid migration) related to the emplacement of the Sebago pluton. The effects of the latter intrusion are widespread in Western Maine (as this fieldtrip demonstrates) with temperatures reaching in excess of 500°C (see below). Furthermore, the abundant pegmatites in this area may also have played an important role, although whether these are temporally related to the Sebago is at present not conclusively demonstrated. The extent to which post-crystallization processes could have produced the mineralogical variation evident in the Songo is debatable. However, it is interesting to suggest a process whereby migration of H_2O through those parts of the Songo adjacent to the Sebago produced localized reducing conditions. As a result Fe_2O_3 contents of the biotites may have been lowered. This in effect would increase the role of titanium with the resultant change to red biotites. Likewise, resorption of sphene, releasing TiO_2 could conceivably result in the formation of ilmenite. Thus many of the mineralogical variations described above may have been produced by such a process. Indeed any fluids eminating from the peraluminous Sebago magma would probably be saturated with regard to Al. Interaction with the Songo might well have caused the peraluminous affinity of those granodiorites proximal to the Sebago.

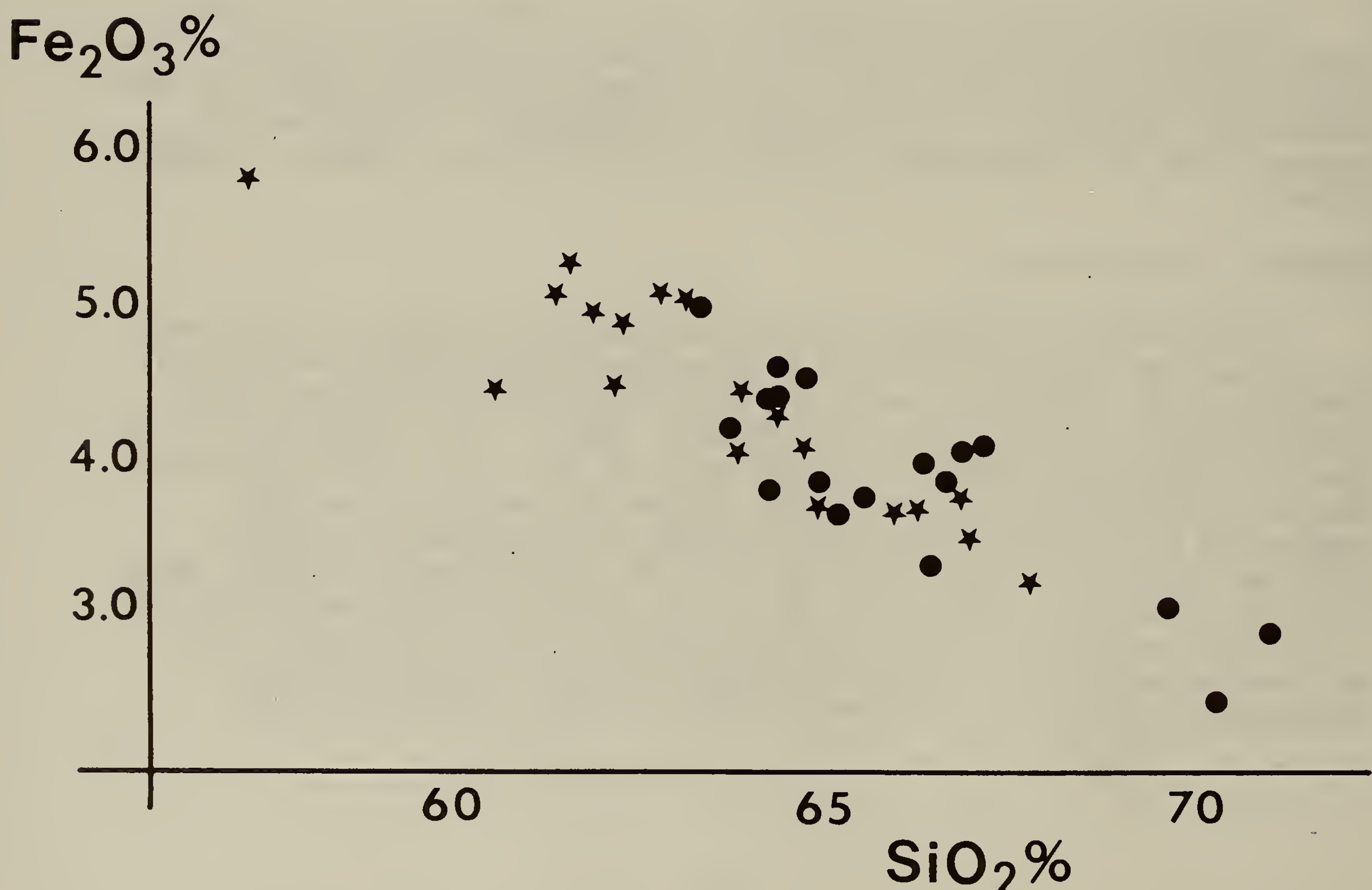
At present this is just speculation and further work is required especially on the mineral chemistry. However, it might well be that the mineralogical variations evident in the Songo are a direct result of the metamorphic effects of the intrusion of the Sebago. In contrast it appears that the igneous geochemistry of the songo has remained largely intact.

Geometry of the Plutons

Since the first regional gravity survey of the Maine, it has been recognized that plutons in the high-grade metamorphic terrane have very little gravity signature. Hence, they have been interpreted as being relatively thin bodies (Kane and Bromery, 1968). The difference in the gravity signature between those plutons that are found in the high-grade metamorphic terrane and those found further to the north has been emphasized by both Kane and Bromery (1968) and Hodge et al. (1982). Detailed gravity surveys across a number of the plutons of western Maine (Carnese, 1983) show them to be thin tabular masses that homotropically dip to the northeast at about 4-6 degrees and extend far beyond their surface contacts. For example, although the Mooselookmeguntic and Reddington plutons are separated on the surface by about 10-15 km.,

Figure 5. Bivariate (Harker) diagrams of TiO_2 and Fe_2O_3(T) wt.% versus SiO_2 wt.% illustrating the compositional variation observed in the Songo pluton. Symbols as for Figure 4.





interpretation of gravity measurements suggests that the Mooselookmeguntic pluton is about 2 km thick and actually extends underneath the Reddington pluton staying within about one km of the surface in the intervening region. Hodge et al. (1982) reached a similar conclusion for the Sebago Batholith and suggested that although it has a surface area of 2600km^2 its thickness is on the order of only 1 km. It is reasonable to assume that it also may extend beyond its surface contacts to the northeast.

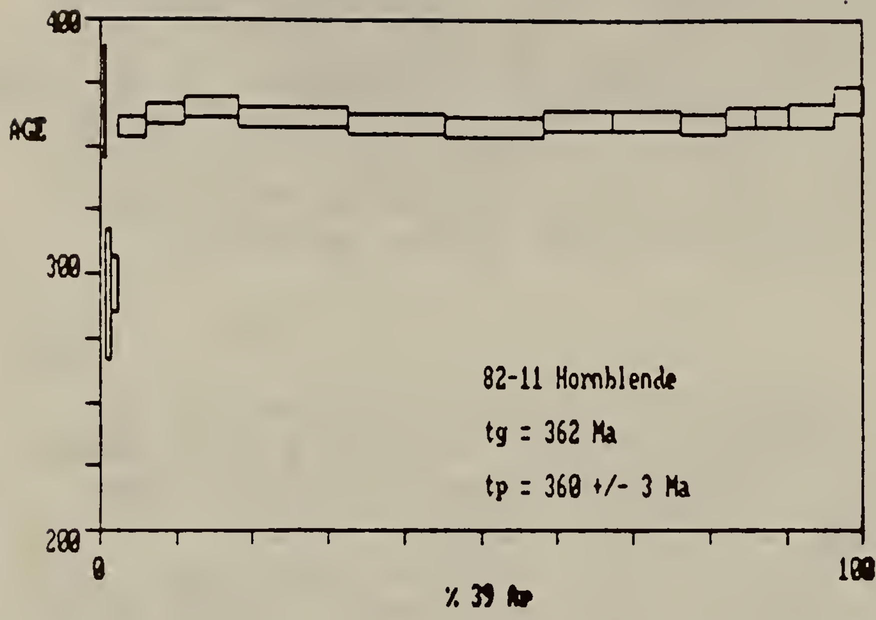
Geochronology

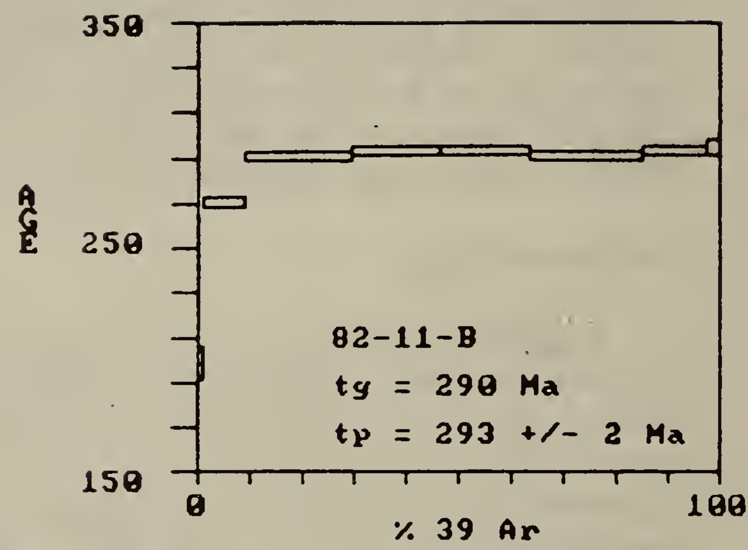
Emplacement ages are known for each of the three main plutons to be visited on the fieldtrip. Moench and Zartman (1976) reported a nine point Rb-Sr isochron age of 371 +/- 6 Ma for the Mooselookmeguntic Pluton that included one point from an aplite dike cutting the Rumford Pluton (corrected to presently accepted decay constants). Samples represent two-mica granites and aplites. In addition, three whole-rock samples from the Whitecap Mountain Pegmatite gave model dates of 350 +/- 6, 356 +/- 6 and 380 +/- 6 Ma, where an initial ratio of .706 was assumed. Interestingly, the three pegmatite samples are roughly colinear and suggest an age of 337 Ma with an initial ratio of 0.737. These dates are consistent with field relationships that indicate the pegmatites are younger than the two-mica granite. Field relationships also indicate that the two mica granite is younger than the hornblende granodiorite and therefore the granodiorite must be older than 371 Ma.

The age of the Sebago Batholith is well defined. Aleinikoff et al. (1985) presented U-Pb data for zircons from two samples collected at different localities. Though distinctly different chords were defined by the two samples both gave upper intercept ages of 324 Ma. Independently, Hayward and Gaudette (1984) used the Rb-Sr whole-rock isochron method and determined an age of 325 Ma for the pluton.

The Songo Pluton appears to be about 380 Ma old, though a precise age has not yet been determined. Four zircon fractions from a single sample (field trip Locality 8) loosely constrain discordia intercepts of 0 and 380 Ma that were interpreted as the result of recent lead loss and crystallization 380 Ma ago (Lux and Aleinikoff, 1985). The samples are not precisely colinear and the cause of this anomalous behavior is unknown at present but may be related to a thermal disturbance at the time of intrusion of the Sebago batholith. The sample is distinctly metaluminous and based on the Zr solubility relationship of Watson and Harrison (1983) and major and trace element data for the sample, it is unlikely that the disturbance is related to the presence of inherited zircons. Rb-Sr whole-rock isochron dating is currently underway which we hope will define a more precise age for the Songo Pluton.

A few representative $^{40}Ar/^{39}Ar$ ages for minerals from a number of the field trip Localities are presented in Fig. 6. With the exception of several hornblendes from the northern part of the Mooselookmeguntic Pluton, mineral ages are younger than the crystallization age of the plutons (Lux 1985; unpublished data). Nonetheless, mineral ages can provide useful information about the ambient temperature in the vicinity of the plutons at the time of intrusion, heating of rocks surrounding plutons, and about the regional cooling





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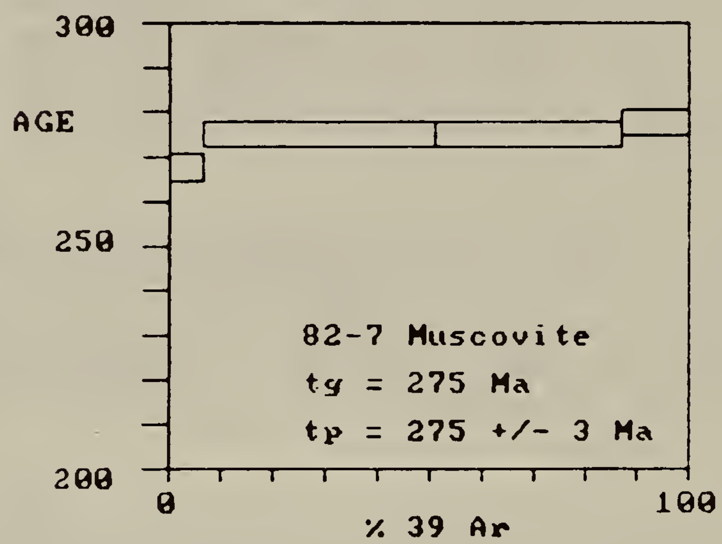

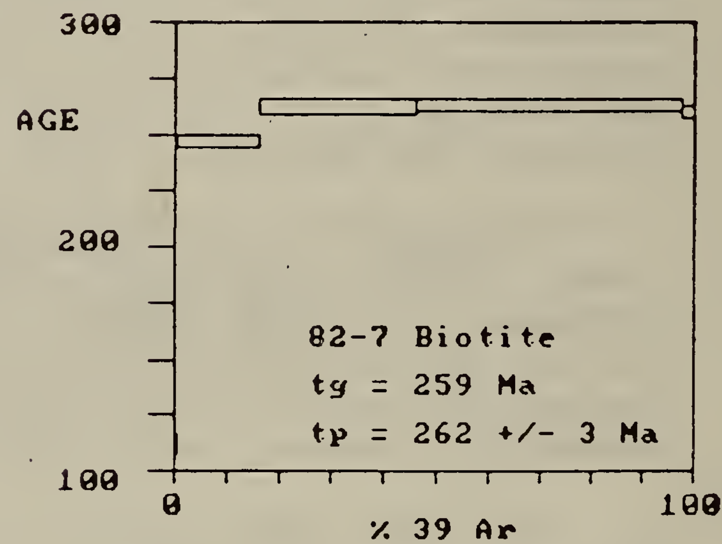

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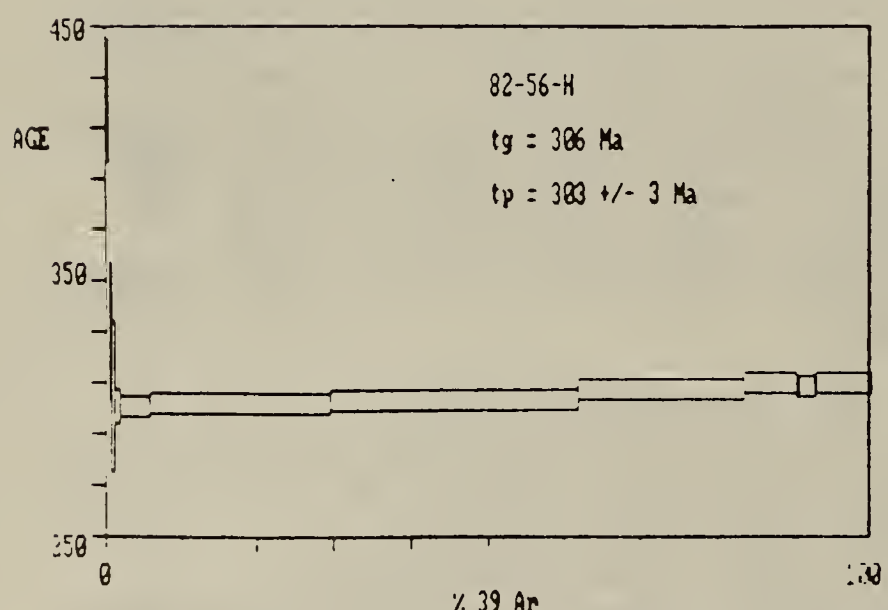

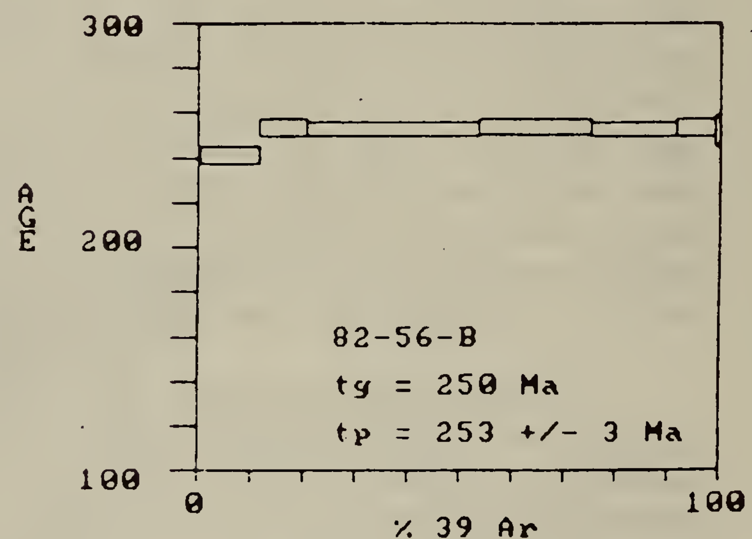

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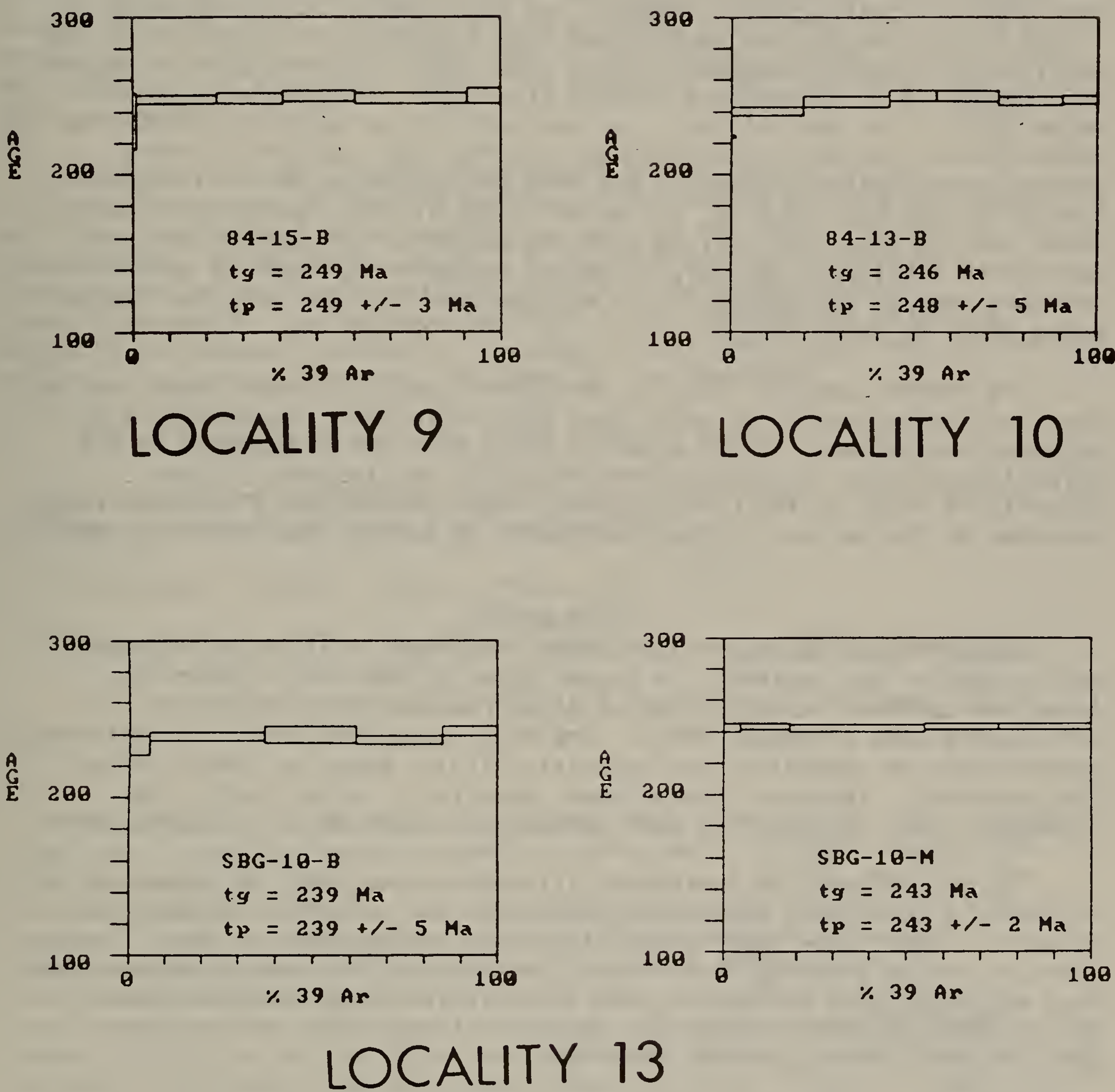

Figure 6. $^{40}Ar/^{39}Ar$ incremental heating release spectra for hornblende, biotite and muscovite. Field trip localities where the samples were collected are given below the spectra.

(depth of burial) of the terrane.

In the context of the Carboniferous metamorphism considered herein, it is the hornblende ages that are of greatest interest. As noted above, hornblendes from the northern portions of the Mooselookmeguntic Pluton have mineral ages coinciding with the crystallization ages. Hence, the rocks must have cooled below 500°C very quickly and experienced no subsequent heating. Moreover, the ambient T of the surrounding rocks must have been well below 500°C. However, as described by Lux and Guidotti (1985), to the south, two hornblendes (one from Locality 3 and one further SW) from the Mooselookmeguntic Pluton show disturbed spectra which suggest some reheating or disturbance at some time after they initially cooled below 500°C (i.e. post 371 MA). In contrast, three hornblendes from the Songo Pluton show much less disturbed spectra that indicate a strong post-crystallization re-heating and then cooling below 500°C at 308.5 ± 1.3, 309.9 ± 1.8, and 305.4 ± 0.9 Ma. As noted in the introduction of this report, these data form an important part of the evidence that much of the area considered herein was strongly re-heated (re-metamorphosed) at essentially the same time as the Sebago Pluton was emplaced (325 Ma) and then cooled below 500°C by about 308 Ma.

The release spectra (Fig. 6) are identified by a sample number and by the field trip locality at which they were collected. The spectra are arranged sequentially in the order in which they are encountered on the trip, i.e. N to S. In this context they will be discussed at each locality in terms of their implications regarding ambient T's of the rocks intruded by the various plutons and depths of burial (regional cooling).

METAMORPHISM

Carboniferous metamorphism reaches K-feldspar + sillimanite grade around much of the northern rim (upper side) of the Sebago batholith. Along the southern (under) side the grade reaches only the upper sillimanite zone (Thomson, 1986). One might speculate that this reflects greater ease of pegmatites and associated fluids escaping upward rather than downward. Textural observations, especially on the south side, (Thomson, 1986) suggest that this metamorphism was largely a static event.

The transition from the upper sillimanite zone (USZ) to K-feldspar + sillimanite zone (KSZ) was studied and described in detail by Evans and Guidotti (1966). The upper stability limit of muscovite was never exceeded and so assuming that P-total was about 3.5Kb (see below) and that P_{H_2O} was high, they estimated T-max to be in the range somewhat below 650°C. More recently, Cheney and Guidotti (1979), using the exchange geothermometer based upon the reaction:

$$(NaAl_3Si_3O_{10}(OH)_2)\text{Musc} + (SiO_2)\text{Qtz} =$$

$$(NaAlSi_3O_8)\text{Plag} + (Al_2SiO_5)\text{Sill} + H_2O$$

obtained more quantitative estimates (623 -643°C for the KSZ). Based on fluid inclusion studies on these rocks, Burruss (1977) obtained very

similar estimates. In the USZ and KSZ the rocks are so recrystallized that there is little evidence left of any earlier Devonian textural features. However, in the S.W. Rumford quadrangle and S.E. Old Speck Mountain quadrangle (on Puzzle Mountain) the rocks are only lower sillimanite zone (LSZ) (Cheney, 1975) and it is believed that some of the textural features (annealed slip cleavages, partial pseudomorphs of staurolite etc) may be remnants from earlier Devonian metamorphism(s). For example, the LSZ rocks are texturally extremely similar to those seen in the Rangeley area (Guidotti 1970, 1974) where only Devonian metamorphic events have occurred. It should also be noted that these textures are essentially identical with those seen on the eastern slopes of the White Mountains in New Hampshire, (Wall and Guidotti, 1986; and Hatch and Wall, this guidebook).

Possibly the most interesting aspect of the Carboniferous metamorphism in the SW Rumford quadrangle is that it affects the very sulfide-rich Smalls Falls formation as well as the Perry Mountain, Rangeley etc. formations. Hence, it involves cordierite-bearing parageneses in the highly sulfitic Smalls Falls formation and staurolite-bearing parageneses in the less or non-sulfitic units. Guidotti and Cheney (manuscript in preparation) have shown that staurolite breaks down before cordierite by means of the reaction:

(1) St + Musc + Qtz = Sill + Bio + Garn + H_2O

Subsequently, end member Mg-Cordierite breaks down (in a more Mg-rich portion of composition space) by the reaction:

(2) Musc + Mg-Cord = Sill + Phlog + Qtz + H_2O

At still higher T's the KSZ is attained via the AKNaCa reaction:

(3) Qtz + Musc + Sodic-Plag = Sill + Sodic-Ksp + Ca-richer Plag + H_2O

As developed in Guidotti and Cheney (Ms), at higher P, the order of reaction (1) and (2) would probably be reversed. It should be noticed from the order of reaction (2) and (3) (See Fig. I also), that in the muscovitic rocks of this area, cordierite breaks down before the KSZ is reached. At grades exceeding the stability of muscovite, as in Massachusetts, end member Mg-cordierite becomes stable again in K-feldspar-bearing rocks (Robinson, pers.comm) and via the subsequent continuous reaction it becomes Fe-richer. Thus, in muscovite-bearing rocks, cordierite becomes Mg-richer until its demise via reaction (2). In contrast, in higher grade K-feldspar-bearing rocks it comes in again as end member Mg-Cordierite which then becomes Fe-richer as T increases further.

The mineral facies diagrams determined for the Carboniferous metamorphism in the SW and SE portions of the Rumford and Old Speck Mountain quadrangles respectively, and southward through the Bryant Pond quadrangle are shown in Fig. 7. Fig. 8 shows the relevant PT curves and suggested PT path (i.e. metamorphic field gradient) for the Carboniferous

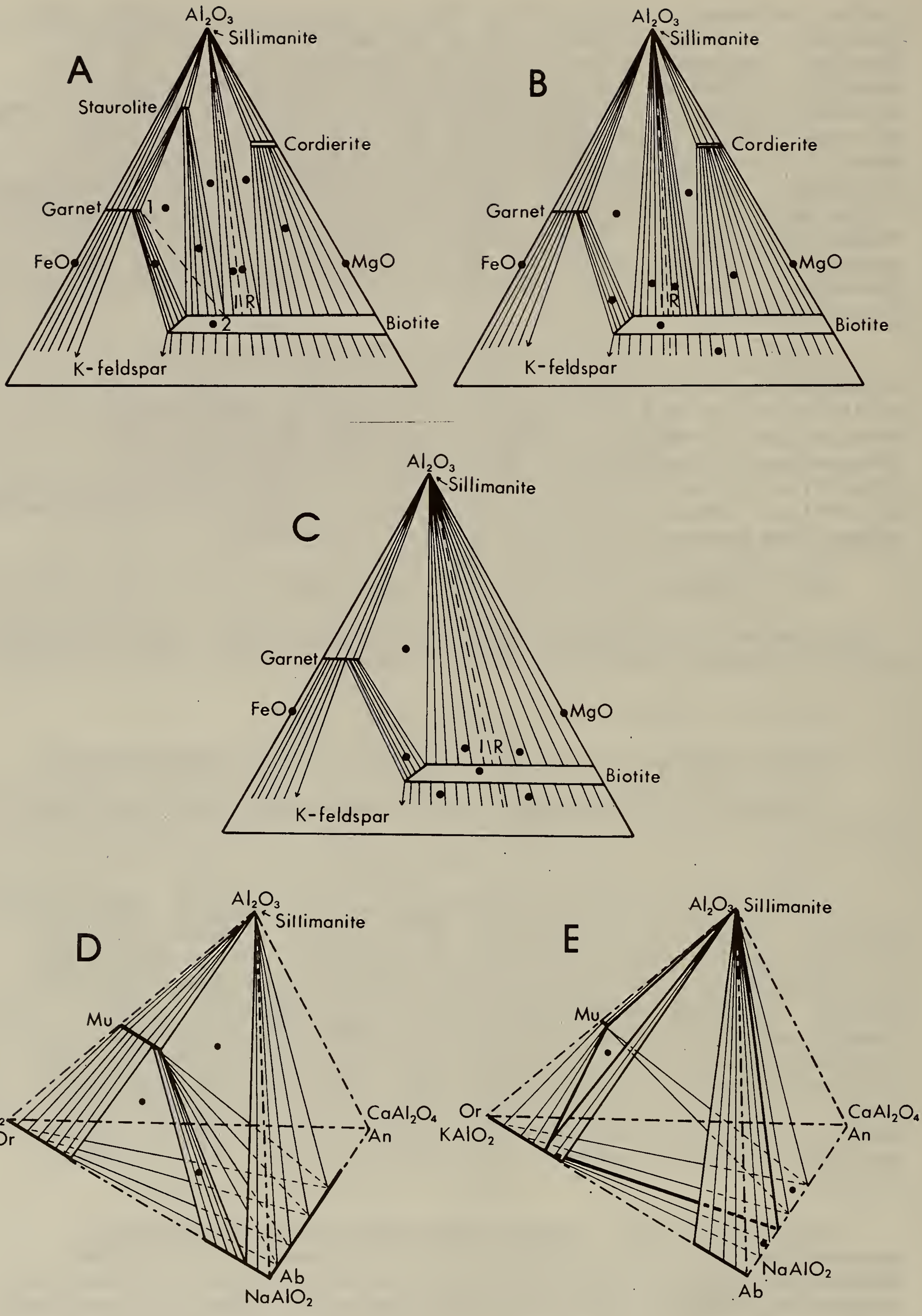
A
Al_2O_3
Sillimanite
Staurolite
Cordierite
Garnet
1
FeO
MgO
R
2
Biotite
K-feldspar
B
Al_2O_3
Sillimanite
Cordierite
Garnet
FeO
MgO
R
Biotite
K-feldspar
C
Al_2O_3
Sillimanite
Garnet
FeO
MgO
R
Biotite
K-feldspar
D
Al_2O_3
Sillimanite
Mu
$KAlO_2$
Or
$CaAl_2O_4$
An
Ab
$NaAlO_2$
E
Al_2O_3
Sillimanite
Mu
Or
$KAlO_2$
$CaAl_2O_4$
An
$NaAlO_2$
Ab

Figure 7. Mineral facies diagrams for S.W. Rumford, S.E. Old Speck Mountain, and Bryant Pond quadrangles. (A)-(C) - AFM projections; quartz, muscovite, and uniform a (H_2O) in all assemblages; I:R indicates boundary between assemblages with ilmenite vs rutile as the Ti-saturating phase. (A) Lower Sillimanite zone; Line 1-2 to indicate garnet in assemblages not shown by AFM projection. (B) Upper sillimanite zone, below the breakdown of cordierite. (C) Upper sillimanite zone and K-feldspar + sillimanite zone, above cordierite break-down. (D) AKNaCa mineral facies diagram for lower and upper sillimanite zones. (E) AKNaCa mineral facies diagram for the K-feldspar + sillimanite zone. Dots-observed assemblages.

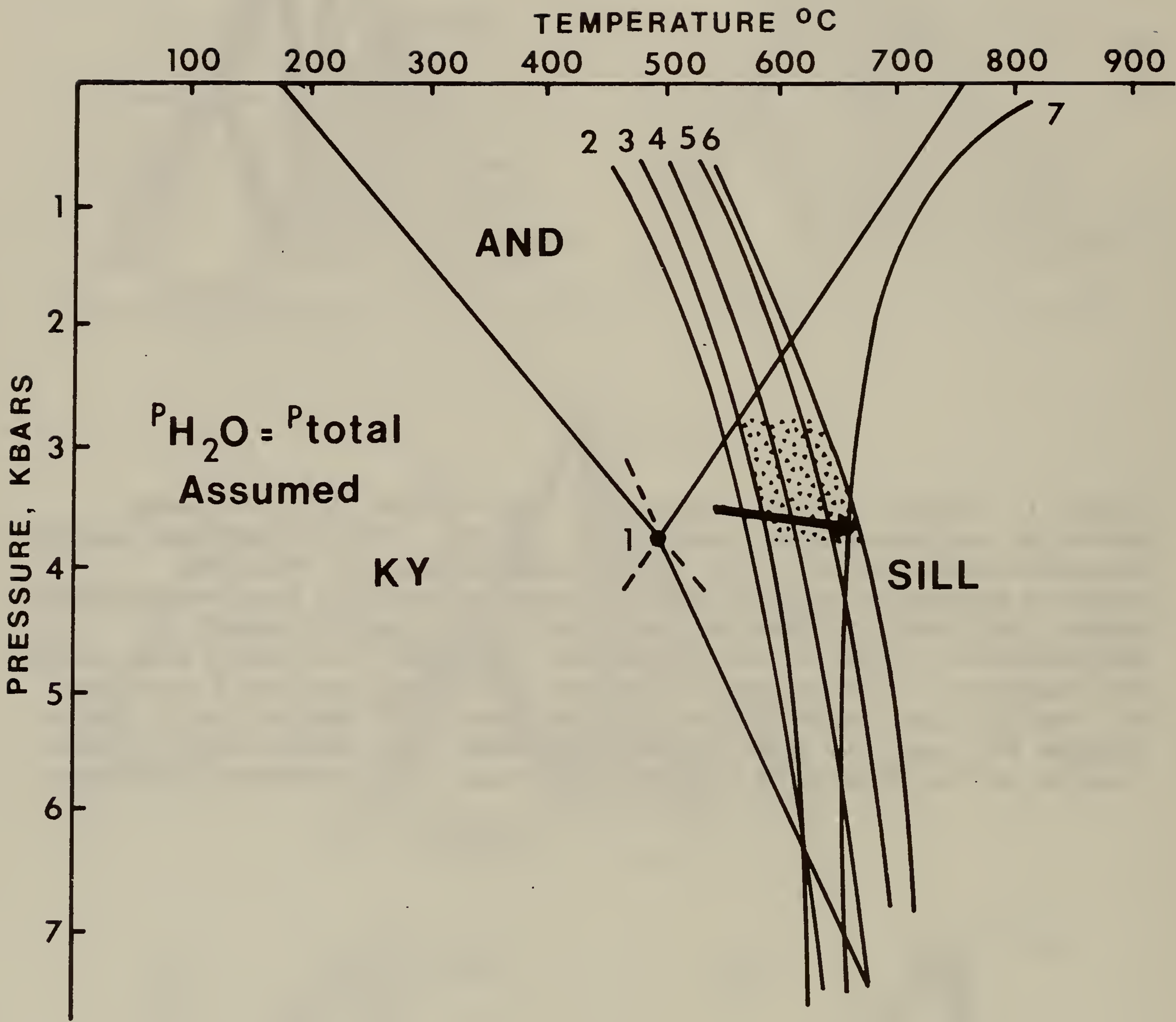

Figure 8. PT grid of equilibria relevant to this report. (1) Al-Silicate curves, Holdaway (1971). (2) Paragonite breakdown, Chatterjee (1972). (3) St + Chl = Al-Sil + Bio, Guidotti (1974). (4) St = Al-Sil + Gn + Bio, Hoschek (1969). (5) Ab + Mu = Al-Sil + Ksp, Chatterjee and Froese (1975). (6) Muscovite breakdown, Chatterjee and Johannes (1974). (7) Granite minimum, Tuttle and Bowen (1958). Stippled area - suggested range of PT conditions for metamorphism considered herein. Heavy arrow - suggested PT path.

metamorphism. It should be noted that in the context of metamorphism due to the thermal effects of the Sebago Batholith (see below) that the PT path is nearly isobaric and because the rocks underwent nearly static recrystallization, the PTt paths of individual horizons were also nearly isobaric.

Another very interesting aspect of the Carboniferous metamorphism in the Bethel and Bryant Pond quadrangles is the effect of the Sebago pluton on the 380 Ma Songo pluton. As noted earlier, the Ar work on the hornblende from the Songo pluton shows that it was re-heated to above 500^0 C sometime after its initial crystallization and cooling, and then cooled again to below 500^0C by 308 Ma. In additon, a number of quite systematic chemical and mineralogic variations were described above for the Songo pluton.

Of particular interest is that these variations are spatially related to the Sebago batholith and its concentration of associated pegmatites. Moreover, several of the variations are exactly those that would result from chemical transfer of material as well as heat from the per-aluminous Sebago pluton into an initially sub-aluminous Songo pluton. Hence, one can at least entertain the suggestion that the Songo pluton is now a metamorphic rock affected by an essentially static heating so that it retains much of its original igneous texture.

Finally, if we consider the effects of the Carboniferous metamorphism to the north in the central and northern portions of the Rumford Quadrangle, we are faced with the more difficult task of detecting the effects of the low grade portions of the metamorphism where it is superimposed on relatively high grade Devonian metamorphism. In the northern part of the Rumford Quadrangle where only Devonian metamorphism has occurred, the rocks are now at staurolite grade (see Locality 1) and involve the assemblage staurolite + biotite + chlorite + garnet. However, it is clear from textural evidence that they were once at higher grade and had the assemblage andalusite + staurolite + biotite + garnet. These two distinct grades would be described to M_3 and M_2 of Guidotti (1970) and Holdaway et al. (1982). Southward, (along the Swift River and Route 17) one finds that the rocks coarsen and that flat lying aplite and pegmatite sills start to occur. Subsequently sillimanite comes in by breaking the staurolite + chlorite join and eventually the rocks become coarse, contorted, migmatitic gneisses as seen at Locality 2.

To date, only modest efforts have been directed at detecting overprinting of the lower grade portions of the Carboniferous metamorphism on the Devonian metamorphism described above. However, several features in part described by Guidotti (1970), clearly reflect a metamorphism post-dating the two Devonian events described above. These features seem best interpreted in terms of the northern effects of the heating caused by the low northerly dipping Sebago batholith. They include:

(I) "In the N and NE sections commonly occur small, euhedral staurolite crystals (lmm) or aggregates of crystals growing on grain boundaries of quartz and plagioclase. In some cases the euhedra occur in the same rock with larger, ragged staurolite grains which contain coarse needles of sillimanite. The euhedral staurolite "appears" fresh and as if

it formed in a later event" (Guidotti, 1970, p. 18).

(2) Further south, in the north central part of the Rumford quadrangle (Roxbury, Locality 2) the Devonian metamorphism was USZ or possibly even KSZ. Now it is at least at garnet grade as garnet and biotite are perfectly fresh but sillimanite is strongly resorbed. For example, sillimanite commonly occurs only as inclusions in quartz and plagioclase, does not cross grain boundaries, and tends to avoid contact with biotite.

(3) Further south in the Rumford quadrangle in the vicinity of Frye and Hale, Guidotti (1970, p. 18) noted: "In the S part of the area, aggregates of fine grained euhedral sillimanite needles occur on quartz and plagioclase grain boundaries. Such aggregates are present in rocks containing coarser sillimanite which is partially resorbed. Again, it "appears" as if the euhedral needles are the result of a late event."

Not enough petrographic work has been done to determine if there is a regular, progressive sequence in the grades established during the Carboniferous metamorphism. In the southernmost Rumford quardangle and into the Bryant Pond quadrangle where the grades are highest there is a simple transition from USZ to KSZ. (Only a very few specimens display any resorbtion textures for the sillimanite.) However, the extensive Devonian dehydration in the central and northern portion of the Rumford Quadrangle may cause complexity in the establishment of diagnostic middle and low grade Carboniferous assemblages. Obviously much additional, detailed work will be required to address this possibility.

THERMAL CONSEQUENCES OF THE INTRUSION OF THE SEBAGO BATHOLITH

As we have discussed, petrologic and geochronologic data suggest that the intrusion of the Sebago batholith was linked to a regional, high-grade metamorphic event. However it may not be immediately clear which is the primary and which is the secondary feature. Here we present a thermal analysis which indicates that the Sebago itself is likely to have provided the heat necessary for the metamorphism.

The Sebago batholith is a laterally extensive sill-like body and, as was concluded above, it is reasonable to assume that the exposed area of the Sebago is a surface expression of a much larger granitic sheet that extends northeast below the present erosion surface and presumably once extended to the southwest above the present erosion surface. With this geometry, the metamorphic terrane to the north of the Sebago lies above its upper surface.

Petrologic observations (Holdaway et al., 1982) of the surrounding rocks indicate that the depth of emplacement of the Sebago was greater than 10km.. Since the thickness is much less than both the width of the batholith and the depth of emplacement, we can approximate the Sebago at the time of the intrusion as an infinite magmatic sheet with uniform temperature, TM, in an infinite solid with uniform temperature, TC. This renders the problem tractable up until the time of solidification. After solidification is completed the thermal evolution can be followed

numerically and the finite thickness of the roof rocks as well as the depth dependence of the country rock temperature can be taken into account.

The important parameters are the specific heat, C, the thermal diffusivity, D, the latent heat, L, the magma temperature TM, and the temperature of the surrounding rock, TC(x), where x is the depth. Varying C, D, and L over the range of reasonable values has little effect on the result. In contrast, variations in TM and TC can have significant effects. Fortunately, studies of plutons of similar composition and depth of emplacement provide reasonable limits on TM. The Strathbogie batholith was found to have had an emplacement temperature of between 750° and 850° C (Phillips et al., 1981) and the Victorian S-type granites of Australia are claimed to have been intruded at 750° to 850°C (Clemens and Wall, 1981; Clemens, 1984). Tewhey (1975) concluded that the Cupsuptic pluton in western Maine had an intrusion temperature of about 800°C. Consequently we have taken TM = 800°C. (Lowering TM to 700°C reduces the maximum attained temperatures near the pluton by about 50°C.)

The initial temperature distribution, TC(x), is much more uncertain. Taking a geotherm inferred by modern day heat flow measurements would no doubt underestimate crustal temperatures since over 10km of crust with its compliment of radioactive elements have been removed since the time of intrusion. Likewise, calculating an equilibrium temperature distribution assuming an extra 10 to 15 km of crust neglects the delay between thickening and warming as well as the effect of erosion and would overestimate TC(x) (see, for example, England and Thompson, 1984). These two procedures would surely give extreme lower and upper limits on TC(x). Instead we have chosen to numerically model the thermal history of the region given geologic and geochronologic constraints on the burial and erosion history. The results shown in Fig. 9, indicate that the Acadian deformation resulted in a significant warming of the crust by the time of the intrusion of the Sebago, but in the absence of the batholith only low grade metamorphism would have occured at depths of 10 to 15 km. The thermal effect of the intrusion, also shown in Fig. 9, is quite dramatic. The modelling predicts a maximum temperature of about 650° C at the upper boundary of the pluton and an array of maximum temperatures in the 2km above the pluton giving an apparent thermal gradient of 80°to 100°C/km. Temperatures in excess of 500°C are predicted up to 2km from the pluton. For a sufficiently extensive sheet dipping at 5°this corresponds to 25 km away from the contact. These predicted conditions are in good agreement with those observed through petrographic and geochronologic studies.

We conclude that the Sebago batholith is likely to have been the source of heat for the surrounding Carboniferous metamorphism, provided that it has a shallow northeastward dip and extends more than 30km beyond its northeastern boundary. If the Sebago is not sufficiently extensive then this model may still be appropriate if, alternatively, the Sebago is one of a number of interconnected sill-like granitic bodies that intruded into the region about 325 Ma ago.

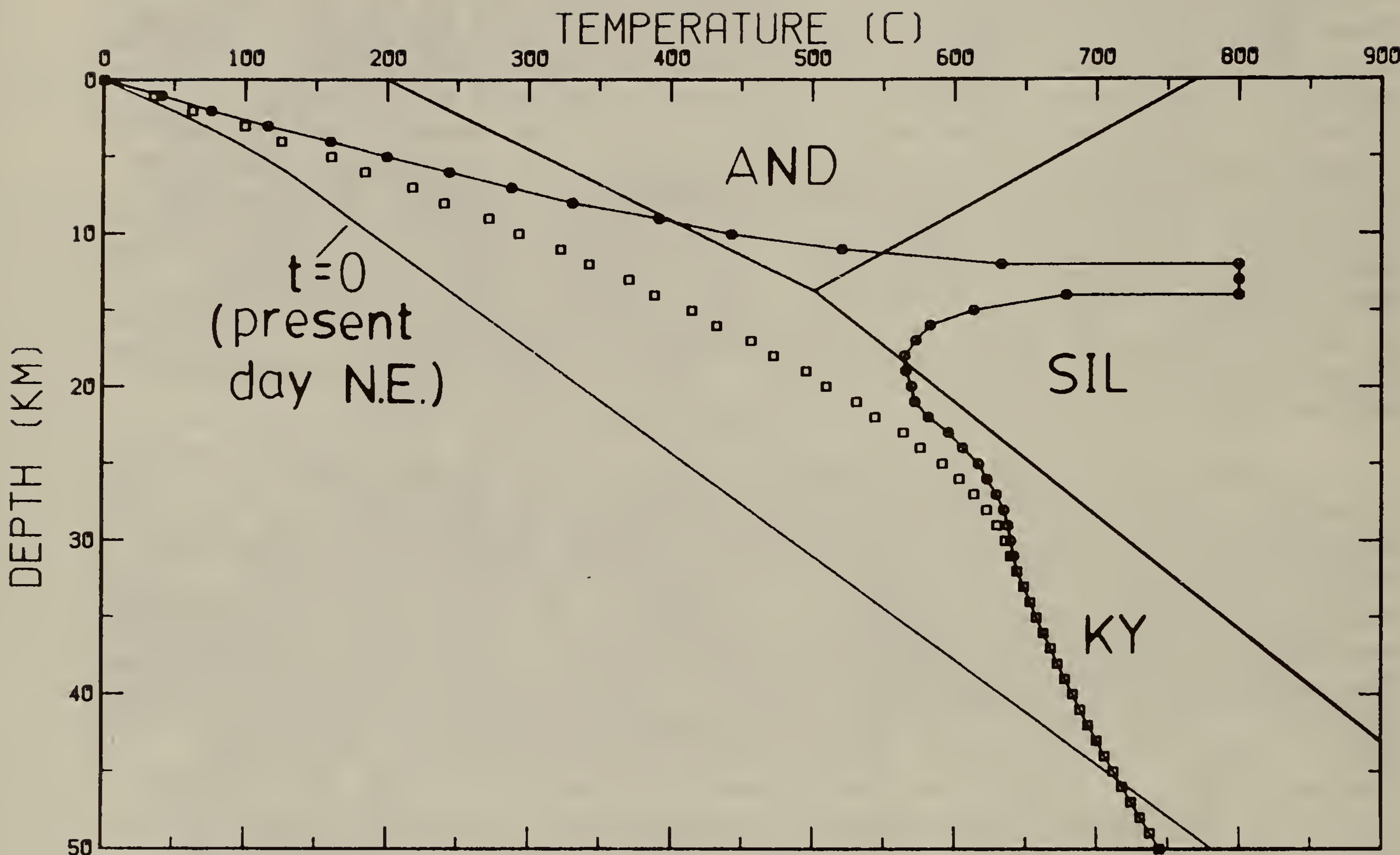

Figure 9. Thermal effect of Acadian thickening and subsequent igneous intrusions. Initial temperature profile, t=0 (440Ma), is that of present day New England. Open boxes: Temperature profile after 115 MY (325Ma). Burial and erosion parameters are: 12.6 km of sedimentation at 0.3 km/MY for 40 MY (440-400 Ma); homogeneous thickening of the sediments from 12.6 to 25 km at 40 MY (400Ma); erosion at 0.05 km/MY from 40 to 115 MY (400 to 325 Ma). Open circles: Array of maximum temperatures near a 2 km thick 800^{0}C infinite magmatic sheet centered at 13km. Burial and erosion history preceding intrusion is the same as for open boxes.

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ITINERARY

The assembly point for this field trip is at the Coos Canyon School in Byron. This site is easy to find as it coincides with the picnic area on the East side of Route 17, about 20 miles north of Rumford. Assembly time is 8:30 AM. Bring lunch materials with you. Topographic maps: Rumford, Bethel, and Bryant Pond 15' quadrangles.

Mileage

0.0 LOCALITY 1 :

Only Devonian metamorphism can be discerned here and the rocks are now at staurolite grade. The diagnostic assemblage is staurolite + biotite + chlorite + garnet + ilmenite + pyrrhotite. This assemblage formed during the M_3 event of Holdaway et al. (1982). In the northern part of the exposure (down on the river edge) staurolite euhedra up to 1 cm in length are found.

Textural evidence for an earlier, higher grade metamorphism (M_2) is seen on bedding-slab surfaces. It consists of coarse muscovite, "Turkey Track" pseudomorphs after andalusite. These pseudomorphs are up to 10X2 cm in size and in a few cases still retain some fresh andalusite. The assemblage formed in this earlier event involved andalusite + staurolite + biotite + garnet.

Note how the pseudomorphs are unoriented within bedding planes. In addition, they are undeformed plus the muscovite in them is unoriented. Moreover, staurolite euhedra and tablets of coarse biotite are also largely unoriented. Although evident mainly from thin section study, these megascopic observations provide some indication that both M_2 and M_3 were largely static events. The rocks here are part of the Perry Mountain fm., (Moench, 1971). Some isoclinal folds are evident on the eastern wall of the gorge and a few scattered aplite dikes are also present in these outcrops. Although these rocks have been affected by two fairly high-grade metamorphisms, it is remarkable how well graded beds and cross-beds are preserved. Possibly this reflects the static nature of the two metamorphisms. Some chemical data is available for the minerals at this outcrop - see specimen 1-8/7/63 of Evans and Guidotti (1966).

For those using this field guide at a later time it would be worthwhile to inspect the rusty - weathering outcrops just to the south on Route 17. These outcrops are unusual for the Perry Mountain

formation in that usually this unit is gray-weathering. Because of the abundant pyrrhotite in these outcrops some sulfide silicate reactions come into play. For example, garnet is absent, staurolite becomes much less abundant or absent and in some of the latter the titanium phase is rutile rather than ilmenite.

The outcrops downstream of the Route 17 bridge are also worth visiting to view nice folds and axial plane cleavages.

Head south on Route 17.

0.2 Bridge over Swift River

4.2 Roxbury post office - Mobil station

5.5 Outcrops of coarse gneiss on left

5.9 More outcrops of the gneiss

6.3 Continued outcrops of gneiss

6.45 <u>LOCALITY 2</u>

River outcrops along Rte 17, south of the village of Roxbury. The rocks here are coarse, swirled, migmatitic gneisses. Bedding is difficult to find. Coarse spangles of muscovite are prominent and in some cases have shapes reminiscent of staurolite crosses. Many spangles contain euhedral garnets similar to those seen in lower grade rocks that retain clear muscovite pseudomorphs after staurolite. Pegmatite veins and irregular masses are prominent throughout this outcrop.

These rocks were originally in the USZ or possibly even KSZ. However, sillimanite is now largely resorbed (not sericitized!) and seen mainly in thin section as inclusions in quartz and feldspar. Hence, the present assemblage is biotite + garnet. Minor chlorite seems to be present only as a very late alteration phase. A few parts of the outcrop and associated loose blocks do retain coarse sillimanite visible to the unaided eye! Can you find it?

Inasmuch as the sillimanite is now largely resorbed but garnet and biotite are fresh, it would appear that these rocks have been affected by a later, medium grade metamorphism. Moreover, because hornblende in plutonic rocks this far north shows disturbed spectra, we would suggest that this later metamorphism reflects the Carboniferous heating event.

The rocks at this outcrop have been mapped by Moench and Hildreth (1976) as undivided Siluro - Devonian.

7.9 Turn to the right on Route 120

8.0 Cross Swift River

8.2 Stop sign - Turn right on Route 120

10.8 Roxbury Notch Summit

13.8 Fork in road - stay left on Route 120

15.4 Bridge over river

15.5 LOCALITY 3 :

Outcrops at Ellis Falls and at roadside along Route 120.

In this part of the Mooselookmeguntic pluton the rock type is a two mica granite which contains abundant inclusions of hornblende + sphene bearing granodiorite (see above). Therefore at this locality two varieties of the intrusion can be examined.

The two mica granite is a fine grained, equigranular rock and is quite leucocratic. Both muscovite and biotite are obvious in hand specimen. In thin section K-feldspar predominates over plagioclase, muscovite is more abundant than biotite and is obviously a primary phase and apatite is the principal accessory.

In contrast to the above are the blocks of granodiorite which are seen in the roadside exposures. The granodiorite is a much darker gray color, has a medium grained, equigranular texture and is distinctly richer in mafic minerals. Biotite and hornblende are abundant, the latter commonly being 6 mm in length. Euhedral sphene is frequently observed in hand specimen.

In many ways the granodiorite, observed here as large inclusions in the two mica granite, is comparable to that which Moench and Hildreth (1976) mapped on the eastern side of the Mooselookmeguntic pluton (see Fig.2).

Release spectra for biotite and hornblende from the granodiorite at this locality are given in Fig. 6 (82-11).

Continue on Route 17.

17.2 Stop sign - turn right, continuing on Route 120

17.5 Road to right is Route 5 north. Stay on Route 120.

17.6 Cross Ellis River

18.05 Downtown Andover - Turn left (south) on Route 5

19.5 View of Whitecap mountain (Pluton) to the south.

23.1 LOCALITY 4 :

Large road cut south of Andover along Route 5.

Present at this locality is the two mica granitic phase of the Mooselookmeguntic pluton. Enclaves of possible metasedimentary origin are also observed along with abundant pegmatites.

The granite here is a light gray, medium grained equigranular rock which is extremely fresh. In places it is slightly porphyritic with larger plagioclase grains. Biotite and muscovite are obvious in hand specimen and significantly small garnet euhedra are also present. In thin section K-feldspar > plagioclase, biotite exceeds muscovite and has the distinctive red-brown pleochroism typical of Ti-rich varieties. Zircon is the dominant accessory phase.

The granite is cut by numerous pegmatitic dikes the largest of

which is Ca. 75cm across, though most are <20 cm. Two sets of pegmatites are apparent at this outcrop, however their relative ages are not clear. They have a simple mineralogy (feldspar, quartz, muscovite) with some garnet present along their edges and internally. In contrast, on Plumbago mt. gem tourmaline has been mined and another large pegmatite body is present on the top of Whitecap mt.

In some parts of this outcrop there are abundant biotite rich enclaves. Some of these appear to have been entrained into the pegmatites whereas others are cut by them. The mineralogy of these enclaves (biotite + feldspar + some garnet) may suggest that they are mafic schists (?) but their origin at present is problematical.

Biotite schlieren are also quite common within the granite. In places there is some evidence to suggest that they are modified basic inclusions but some may simply be biotite rich segregations.

Release spectra for muscovite and biotite from this locality are given in Fig. 6 (82-7).

Continue South on Route 5.

25.2 LOCALITY 5 :

This outcrop involves well-bedded Perry Mountain formation. Graded beds show that the strata are overturned at this locality.

The grade here is LSZ and the typical assemblage involves sillimanite + staurolite + biotite + garnet + ilmenite + pyrrhotite. Staurolite is mostly replaced by coarse muscovite to form the conspicuous white "eyes" seen in the pelitic portion of the graded beds. For the most part the staurolite is seen only in thin section but occasionally it is visible in hand specimen. In some places the muscovite "eyes" approximate the shape of staurolite and contain euhedral garnet inclusions. The rough weathering surface of the pelitic beds is largely due to very abundant clots of fibrolitic sillimanite.

A slip cleavage at high angles to bedding is evident in parts of this outcrop. In thin section it is seen that this slip cleavage has been annealed to form polygonal arcs. Moreover, this cleavage has no effect on the muscovite pseudomorphs after staurolite.

These LSZ rocks trace directly up grade to the USZ and then SKZ and so it is believed that the last equilibration occurred during the Carboniferous heating event.

25.7 Leave Route 5, turn left and cross Ellis River

26.1 Bear right

27.5 Continue on the main road

30.2 LOCALITY 6 :

Just north of Rumford Center.

This outcrop consists of coarse mica schist and is interbedded on a thick scale with biotite granulite. Moench and Hildreth (1976) map this rock as part of the Perry Mountain formation.

The grade here is USZ (Carboniferous age) and only indestinct eyes of coarse muscovite persist. Fibrolitic sillimanite is abundant on biotite plates and some sheath-like sillimanite covered surfaces occur. The main assemblage is sillimanite + garnet + biotite + ilmenite + pyrrhotite.

A good lineation is present due to a finely spaces slip cleavage which intersects bedding at a high angle.

30.6 Route 2 is encountered - bear right

31.5 Outcrop of Perry Mountain Fm.

33.6 Rumford Animal Park

34.4 Rumford Point - turn left on Route 232 and cross Androscoggin River

34.9 Bear left on Route 232

37.0 Bear right, staying on Route 232

40.8 Crossing through an esker

41.1 Riding on the esker

43.6 Outcrop of Songo granodiorite

43.75 Stop sign - junction with Route 26; bear left.

44.9 Outcrop of Songo on the left

45.7 More Songo

47.5 <u>LOCALITY 7</u> :

These rocks are in the SKZ. Sillimanite is extremely abundant and K-feldspar is abundant both in the groundmass and as coarse megacrysts. The latter commonly contain quartz inclusions that simulate bipyramids such as one would see in volcanic rocks. Muscovite is still stable so that the full assemblage is sillimanite + garnet + biotite + K-feldspar (orthoclase) + muscovite + plagioclase (An_{30}) + ilmenite + pyrrhotite. This outcrop is part of the extensive SKZ that surrounds the Sebago batholith on its northern side.

47.8 Intersection with small cross road - carefully make a U-turn and head back up Route 26.

51.9 Pass the junction with Route 232, continue on Route 26.

52.1 Outcrop of Songo

53.1 Downtown Bryant Pond, post office - bear right, staying on Route 26

53.5 Outcrop of Songo

55.25 Outcrop of Songo

55.45 View to the right, across lake - Bucks Ledge, an exfoliation dome in Songo Pluton

56.3 Locke Mills - Songo outcrop

57.0 Bethel Town Line

60.1 Telstar High School

61.4 Entering town of Bethel

61.5 Cross railroad track - leave Route 26 and continue straight on Route 35 - Main street of Bethel

61.7 Continue up Main street of town - <u>do not</u> follow Route 35

61.95 Stop sign at top of hill. Bear left to village green and immediately bear right on Route 5. Bethel fire Station in view. Stay on Route 5 turning right at the Opera House Condominiums.

62.15 Stop sign where Route 5 takes sharp left; proceed <u>straight</u> across onto minor forest road.

64.25 <u>LOCALITY 8</u> :

Small ridge outcrop approximately 25 yds off the forest road, NE of Sparrowhawk mt. This outcrop consists of the biotite + hornblende + sphene variety of the Songo pluton. The granodiorite is a medium grained, relatively mafic rich rock which contains hornblende + biotite with sphene as the dominant accessory phase. Both hornblende and sphene are obvious in hand specimen as is the distinctive black coloration of the biotites (cf. to Locality 9). In thin section plagioclase greatly predominates over the commonly anhedral (interstitial) K-feldspar, biotite has 'normal' brown/green pleochroism and apatite and zircon are also present. The granodiorite at this outcrop is foliated, this being defined by the biotite, hornblende and plagioclase grains. However in places the deformation is more intense and small scale folding is observed. In general the texture and mineralogy evident in the granodiorite here is typical of most of the NW, N and NE parts of the Songo pluton. (see Fig. 3). However in some areas hornblende is not present and the granodiorites are not as deformed as is apparent at this

outcrop. Release spectra for hornblende and biotite from this locality are given in Fig. 6 (82-56).

66.7 Sharp right turn in the road

67.5 Intersection of forest roads. Take a left turn and proceed south.

69.7 <u>LOCALITY 9</u> :

A series of small pavement outcrops alongside the forest road at the abandoned schoolhouse.

In travelling further south into the central zone of the Songo pluton, the Ti-rich biotite granodiorites are observed along with well developed deformation features.

At this locality the granodiorite is essentially medium grained but is somewhat porphyritic with larger plagioclase grains. Biotite is the only mafic phase and on close examination it has a distint red - brown sheen, in contrast to those observed at Locality 8. The red - brown pleochroism is more clearly observed in thin section. Other main differences to the previous stop are a) the absence of hornblende and b) a lack of sphene, with apatite now being the dominant accessory.

The granodiorite at these outcrops displays an intense foliation this trending generally N - S and orientated along this are large plagioclase grains some of which are up to 3 cm in length. In some places incipient mafic and felsic banding is observed.

A release spectrum for biotite from this locality is given in Fig. 6 (84-13).

70.0 Take road to Crocker Pond Campground (sign posted)

71.5 Crocker Pond campground. Park at sign for campground 5 close to the restroom (!). Proceed to this campsite and then take trail which goes south along lakeshore for approximately 50 yds.

<u>LOCALITY 10</u> :

Lakeside outcrops of the Songo granodiorite at the SE end of Crocker Pond.

The granodiorite is medium grained and strongly foliated and deformed which is typical for this general area of the pluton. The red - brown color of the biotites is again observed and there is also some sparse muscovite present.

Of note at this outcrop is the intense foliation and possible folding. Some of the banding is comparable to that seen at the previous locality.

Also observed here is a large metasedimentary xenolith which is possibly a block of the local country rock (Madrid fm.?). These are not particularly common but are present in this area. They do not appear to have interacted to any large extent with the granodiorite.

Proceed back along the road to main forest road.

73.2 Turn right at end of Crocker Pond road.

76.2 Junction with Route 5. Turn right and proceed South.

78.1 Junction with Route 35, keep on Route 5.

79.0 Pass Bumpus Mine pegmatite quarry on Route 5.

80.5 LOCALITY 11 :

Large roadcut along Route 5.

This outcrop exhibits an association of rock types which is frequently observed at many outcrops in this general area of the Songo pluton. The granodiorite is cut by a large pegmatite body and a two mica, garnet-bearing granite is seen within the latter.

The darker, foliated/banded Songo granodiorite is easily recognized. It again contains Ti-rich biotite, no hornblende and very rare sphene. In thin section some muscovite is present and apatite is abundant. The foliation observed trends NE dipping at 35 to NW.

The granodiorite is cut by a large two mica, feldspar, quartz pegmatite. In some places blocks of Songo granodiorite are observed within the pegmatite and these are seen to be strongly deformed.

The two mica garnetiferous granite is seen in close association with the pegmatite. It is a fine to medium grained leucocratic rock with obvious muscovite and garnet clots. Its field relations to the pegmatite are ambiguous - in some cases it appears to have cut the pegmatite whereas in others this is not so clear pointing to synpegmatite emplacement.

In this part of the Songo, the close association of pegmatite and two mica garnet granite is a frequent occurrence. The origin of the latter rock is problematical and depends to a large extent on the age of pegmatite emplacement.

This outcrop may have been faulted as is suggested by the large vertical face in the pegmatite on the E side of Route 5. Some limited mineralisation is observed along this face.

A release spectrum for biotite from this locality is given in Fig. 6 (84-15).

84.3 Junction of Routes 35 and 5. Turn left at the signpost and proceed east along Route 35.

85.2 Fork in road at North Waterford - stay left which is Route 118.

86.4 LOCALITY 12 :

Small roadside outcrop along Route 118 and in the adjacent Crooked River.

At this locality the two mica - Sebago granite is observed at one of its closest points to the Songo pluton.

The Sebago granite here is a medium-grained, equigranular rock

with an overall light brown color due largely to the K-feldspars. Biotite and muscovite are obvious in hand specimen and present in roughly equal amounts. However, in other parts of the pluton muscovite predominates.

Proceed to the east on Route 118.

90.5 Junction of Routes 118 and 37. Keep going east on Route 118.

93.9 LOCALITY 13 :

Roadside outcrop of the Sebago granite at Little Penessawassee Pond.

The Sebago granite observed here and at Locality 12 are fairly typical of this N part of the pluton.

Again the rock is medium grained and equigranular with obvious biotite and muscovite. In thin section the biotites display a distinct red-brown pleochroism (cf. to some parts of the Songo) and exceed muscovite in abundance. K-feldspar is dominant and apatite, zircon and some opaque oxides are present as the accessory phases.

Release spectra for biotite and muscovite from this locality are given in Fig. 6 (SBG-10).

96.7 LOCALITY 14 :

Discussion of Thermal Modelling of the Sebago pluton

For those heading south, continue east on Route 118 to junction with Route 26. Turn right on 26 and it will take you to the Maine Pike. Before heading down Route 26, you may want to note the symbol in the lower, right-hand corner of Fig. 1.

TRIP C-5

SEAWALL BEACH AN UNSPOILED BARRIER SPIT

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INTRODUCTION

Seawall Beach "is the only large, undeveloped or unaltered barrier spit in Maine." (Nelson and Fink 1978) Most barrier spit systems in Maine have been developed or altered by man for his uses. The system is typically connected to the mainland by a rocky headland and extends northeasterly to southwesterly by a sand beach and dune systems. It terminates on the southwest by the Sprague River and is separated from Popham beach on the northeast by the Morse River. Both the Sprague River and the Morse River are tidal inlets. The Sprague River meanders from its source through an extensive tidal marsh system. Between the beach and the marsh on the southeast end is a parabolic dune system which rises at its highest to about 4.5 meters above the high tide level. On the northeast end of the barrier the marsh system is not as well developed and the Morse River terminates in a tidal embayment called Spirit Pond. The dune system on the north is relatively low in elevation and has in the past developed a low scarp as erosion primarily from northeast storms have impinged on it. This dune system is also very much influenced by the position of the outer part of the Morse River Channel which has over the years migrated across the segment between Morse Mountain and Popham Beach and at times infringed upon the dunes.

Seawall Beach is a part of the nature preserve called Morse Mountain under the care of Bates College. In its unaltered state it provides an area of research which can be carried through over a number of years,giving long range information. Both the Geology Department and the Biology Department have a number of ongoing projects and Seawall has provided research areas for a number of senior theses. Many courses give students their first opportunity to view the action of the seashore environment at this location.

The purpose of this trip is to give the opportunity to view the characteristics of this unique area. Many if not most of the ideas expressed here are those of a number of persons who have been researching the area. Acknowledgement is given Bruce Nelson,Ken Fink,Duncan Fitzgerald and his students. I am also indebted to my students Carol Gerster,Ross Hanson,Laura Simmons, Amy Frankenburg who did theses there,also I thank my students in

my Short Term Course of 1986, Keith Wight, Jennifer Andrews, Jennifer Smalley and Steve Browning who aided me with specific data collection for this trip. The research thus far accomplished still only gives suggestion of the "facts" but hopefully with long term results we will know more.

Only very basic facts are presented in the field log, it is hoped that persons will be stimulated to research some of the areas untouched and that discussion will be initiated. During the trip graphics will be displayed to illustrate many of the features observed.

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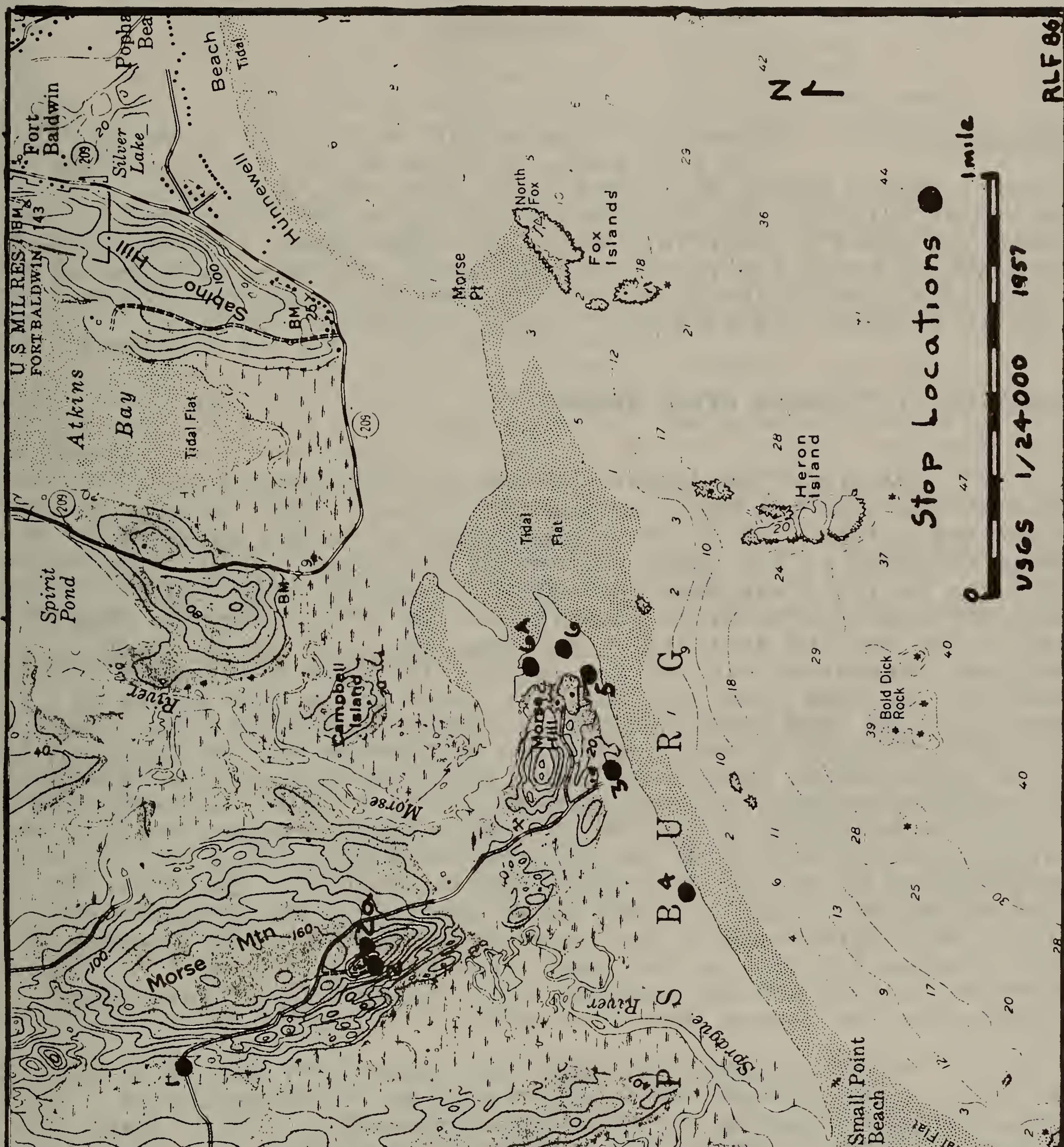
U.S. MIL. RES. FORT BALDWIN
Fort Baldwin
Silver Lake
Hunnewell Beach
Sabino Hill
Atkins Bay
Tidal Flat
Spirit Pond
Morse River
Campbell Island
Morse Mtn
Morse Hill
Morse Pt
North Fox
Fox Islands
Heron Island
Bold Dick Rock
PHIPPSBURG
Sprague River
Small Point Beach
Small Point
Small Point Hill
SMALL POINT ROAD
Small Harbor
Cape Small
N
1 mile
Stop Locations
USGS 1/24000 1957
RLF 86

Fig. 1

ITINERARY

Assembly point - McDonalds, Topsham Fair Mall (I-95 & Rte 196) 7:30 A.M. Route to Morse Mountain Preserve and Seawall Beach: Follow Route 196 to Route 1. North on Route 1 to Bath. At Bath turn south on Route 209, at about 9.5-10 miles Route 209 takes a sharp left to go to Popham Beach. Continue straight on Route 216 for about one mile. Opposite mail box on right is a road to the left. Parking area before gate. Follow road to causeway and Location 1 about .5 miles.

LOCATION 1: SPRAGUE RIVER MARSH

We are at the upper portion of the Sprague River. Here at low tide the small stream falls over a rocky bottom under the bridge. It is often questioned as to how much fresh water is added to the upper reaches of the river. On May 22, 1986 the salinity was measured with an incoming tide. The salinity was 19.5 0/00,the temperature was 9^{0} C.. When the tide was at low ebb the salinity at the same location was 10 0/00 and the temperature was 16^{0} C.. During high tide as it was reaching its peak here the salinity below the bridge showed at about 22 0/00 and above the bridge rapidly increased from about 8 0/00 to 22 0/00 as the tide passed up the falls. Two days later measurements again were made, this time on the upper side of the bridge where the temperature was 8^{0} C. with a salinity of 7 0/00 and below the bridge still under tidal influence the salinity was 18 0/00 with the same temperature. These measurements were made on a cloudy day during which there was a limited amount of rain. Under sunny conditions the temperature will rise considerably in that portion above the bridge as the tide changes.Salinity on the beach that day was 26 0/00. Although this is but one example, it is evident that fresh water is entering the system above the bridge .

The conclusions are that the amount of fresh water in the tidal inlet depends on several factors. 1. The amount of precipitation immediately prior to measurement. 2. The state of tide. 3. The location within the inlet system.

LOCATION 2: LOOKOUT POINT ON TOP OF MORSE MOUNTAIN

From here one can look out across Sprague River Marsh. The regular course of the river is marked by the meandering pattern. Some years ago a straight ditch was made in order that the marsh drain more rapidly in an attempt to control the mosquito population. On the marsh can be seen a number of small

ponds called pannes. The origin of these is not yet known. In some cases they seem to be a wide place in a small drainage channel, however, in most cases they are not connected to the tidal creek. Some suggestions have been made that they are rot spots in the vegetation or perhaps where ice rafted sediments are dropped killing the vegetation. They are a good research project.

Looking to the south one can see the south end of Seawall beach. At the extreme right against the steep banking on bedrock where the cottages are located you can see where the Sprague River enters the sea. If it were not for the river the beach would be tied to the mainland here. On the beach side a spit has developed behind which are located a dune field, which is bordered on the back side by the marsh.

Looking back to the left about a third of the way along the beach you can see a large mound with a depression. This is probably the most spectacular of the parabolic dunes to be observed on this beach. As you can see the prevailing wind is from the the northwest and blows the sand leaving a blowout and carrying the sand toward the ocean. On the oceanward side you can observe the sand which has been transported and accumulates over the sea grasses on the ocean side. The dunes are cut into by the waves in storms particularly in the northeast storms of the winter gradually cutting away the dune.

LOCATION 2A: MORSE MOUNTAIN VIEW

Walk over to the north side of the mountain on the bare rock slope and we can look onto the tidal inlet named the Morse River which separates the Morse Mountain Preserve from Popham Beach. The Morse River is much larger than the Sprague and ends in the body of water to the left called Spirit Pond.. Near the river you can see a bedrock outcrop of granite pegmatite which prevents lateral migration of the Morse River. At the main beach we can also see such an outcrop which has tied the beach to the headland. We will look at this in more detail at Location 6.

LOCATION 3: THE BEACH

We have arrived at the beach itself. On the left is a rock outcrop of granite pegmatite. An examination of this outcrop shows the intrusion of quartz veins, much biotite mica and hornblende, as well as occasional inclusions of country rock.This leads us to a discussion of the source of sediment on the beach. During the spring of 1986 a series of samples were taken on three segments of the beach to see if there are any suggestions of variations either in size or composition along the length of the beach. Several samples were taken from the spit at the high tide line near the Sprague River. The three segments mentioned were taken as follows: one toward the south

end of the beach, one in the middle sector near the pegmatite outcrop and the third on the north end of the beach. The system involved making three traverses 30 meters apart and perpendicular to the length of the beach. On each traverse samples were taken at 30 meter intervals from high to low tide lines. Fitzgerald and Fink (1980) give the source of sediments as related to the Kennebec River, from glacial deposits,local bedrock and biogenic sources. They also present a scenario of a clockwise gyre recirculating the sands from the Kennebec outward and southward and then back to the north along the front of Seawall and Popham beaches.

The general conclusions are that a fining occurs north along the beach, indicating probable net drift in this direction. Dye marker experiments show the movement of water along the beach to be variable, very much dependent on wind direction.

LOCATION 4 : PARABOLIC DUNE

Traverse a distance of 1200 meters south along the beach. The dune line up to this point has a relatively low profile. Here the dunes increase to a height of about 4.5 meters above the high tide line. During May of 1986 one of the most noticeable features seen was the washover into the blowout area behind the dune. The strongest winds particularly in the fall and with clearing weather after a storm are from the northwest and hence most modification of the dunes occurs with these winds. Sand is blown out from behind the dune and across the crest. In the lee of the dune (in this case on the oceanward side), the sand is deposited on the dune grass. The seaward side is further modified by winds coming from offshore, and probably more importantly is eroded by storm waves. The washover you observe is a result of these storms.

LOCATION 5: LOW DUNES ON NORTH END OF BEACH

Returning back up the beach, past the bedrock we are struck by the low dune profile. Here the front edge of the dune very often shows a steep low scarp,but is not very high. Northeast storms are most likely to erode this part of the beach, although the rock islands just off-shore should logically protect them. More than likely the combination of wave action plus the tidal currents out of the Morse River are responsible for most of the erosion. In the past few years this segment of the back shore has been colonized by beach grass, where in 1982 Gerster found this area to have the above mentioned scarp. The channel of the Morse River has shifted toward Popham since that time. Wind action is responsible for the transport of some of this sand to again build up these low dunes. In 1982 one could see up on this portion of the beach similar features now seen further to the northeast on the tidal flat.

Another mechanism of sand transport is that by floating ice blocks in the winter season. At low tide ice blocks are stranded on the tidal flats, with a return of the tides these blocks are picked up and floated shoreward. As they rested on the flats sand is frozen to the bottom side and with melting it is deposited. In the salt marshes one can observe this phenomenon more readily and several persons have suggested that this may be responsible for the "dead" spots on the marsh where salt pannes form.

Looking seaward from here one can observe a series of islands. Heron island is to the southwest and the Fox Islands to the northeast in front of Popham Beach. (See Figure 1) On a day with incoming waves this is an excellent place to observe refraction and reflection of wave fronts. This produces a low tide tombolo because of the wave interference pattern. Observations of ripples on the tidal flat give us the opportunity to see evidence of the multiplicity of wave directions.

LOCATION 6: MORSE RIVER

Here there are several features of note. During the five or so years that we have been observing the beach, it is very noticeable that the beach grass has repopulated the bar, as previously mentioned. In 1982 when Gerster was making her survey there was beach and tidal flat in front of the low scarp of the dune. In that year the channel extended almost straight out across the bar. In the winter of 1982-83 the channel shifted its position roughly to the present location. With tide receding, drainage occurs first across the bar , as can be seen at low tide with the ripple patterns, then becomes confined to the main channel.

The second feature to note is that as you walk along the inlet your feet suddenly sink into the sands. These sands were investigated and determined to have a very narrow range of sizes, hence low stability. This seems to the result of the constant shift between ebb and flow conditions,hence, sorting is excellent and any of the fine sediments are soon transported away. Equigranular sediments are quite unstable.

On the left you can observe a bedrock knob. This is the one we observed from the top of the mountain. Its main effect is to limit the lateral migration of the Morse River as it meanders. Composition of the knob is granite pegmatite. The tidal inlet has not always occupied this position. Several authors notably Fink and Fitzgerald have used areal photographs to show former shoreline features between here and Popham beach. In the far distance where you can observe the Pitch Pine forest is the scarp which represents the location of the tidal inlet in times past.Across the inlet you are also able to observe a break in the low dune where in 1982 a washover occured

and the dune sands were carried into the marsh behind.

LOCATION 6A: TIDAL DELTA

Probably the most significant features of interest in the tidal inlet are the tidal deltas. Hanson (1983) and Frankenburg (1984) in their undergraduate theses studied these deltas in detail. These are best observed during the low tide cycle. These are flood tidal deltas constructed within the tidal inlet channel and found in both the Sprague and Morse Rivers. These deltas show the typical characteristics found in such structures. Megaripples develop with the flood tide and shift direction with the ebb tide. The flood tide comes across a ramp and the ebb tide is diverted with a shield to the other side of the delta.

Frankenburg (1984) has made the most detailed study of the outer delta in the Morse River. She made observations from May of 1983-December 1983. She established a "permanent center stake" and from this each month reconstructed a 25m x 25m grid. "Measurements were made from the outermost stakes at the time of the daily predicted low tides." (Frankenburg 1984) The shape of the lobe was determinined for comparison with the previous month and sand samples of 6 or 7 cm^3 were taken from the ebb shield, the flood ramp, the flood channel, and the spillover lobe for size analysis and comparisons.

In general it is accepted that the flood tide is stronger than the ebb tide. As the tide comes in it is forced into the narrower inlets hence an increase in velocity. Gravity is the governing force with a shift in tide and hence lower velocities are experienced with the ebb. Current velocities are also greatly affected by difference between the spring and neap tides. With a spring tide more water must enter the inlet over the same time span hence greater velocities.

General conclusions by Frankenburg are that the flood-tidal deltas in the Morse River are stable features. "Modification in the form of accretion and depletion of the deltas occurs on a regular basis due to the effects of a changing tidal prism. The sediment size decreases up inlet as distance from source is increased. The effects of weather on the delta is of importance when connected with the effects of tidal currents."
(Frankenburg,1984.p.87)

TRIP C-6

THE STRUCTURAL AND STRATIGRAPHIC DEVELOPMENT OF THE CASCO BAY GROUP AT HARPSWELL NECK, MAINE

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and

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INTRODUCTION

This field trip provides the opportunity to examine in detail the lithologic, textural and structural variations within the Casco Bay Group formations at a single area, Harpswell Neck (Figure 1). Regional stratigraphic and structural relations throughout the Casco Bay are beyond the scope of this field trip guide. The overall structure and lithologies were originally described by Hussey (1971) as part of the geology of the Orrs Island quadrangle. Detailed structural and lithologic mapping within this original framework was conducted by Swanson and Pollock during summer field camp excercises 1983-86.

STRATIGRAPHY

The lithologies within the Harpswell Neck area include representatives from the entire stratigraphic sequence of the Casco Bay Group, exclusive of the Cushing Formation. Table 1 summarizes the representative lithologies and the stratigraphic sequence. The exposed anticlinal structure on Barnes Island is regionally important in establishing the stratigraphic sequence of the Casco Bay Group. All contacts are sharp and conformable. The Cape Elizabeth - Spring Point contact is locally gradational over distances of 30 - 40 cm. The lithologies observed here are typical of these formations elsewhere. However, the exposed thicknesses exhibit wider ranges than other areas in the Casco Bay due to textural and structural

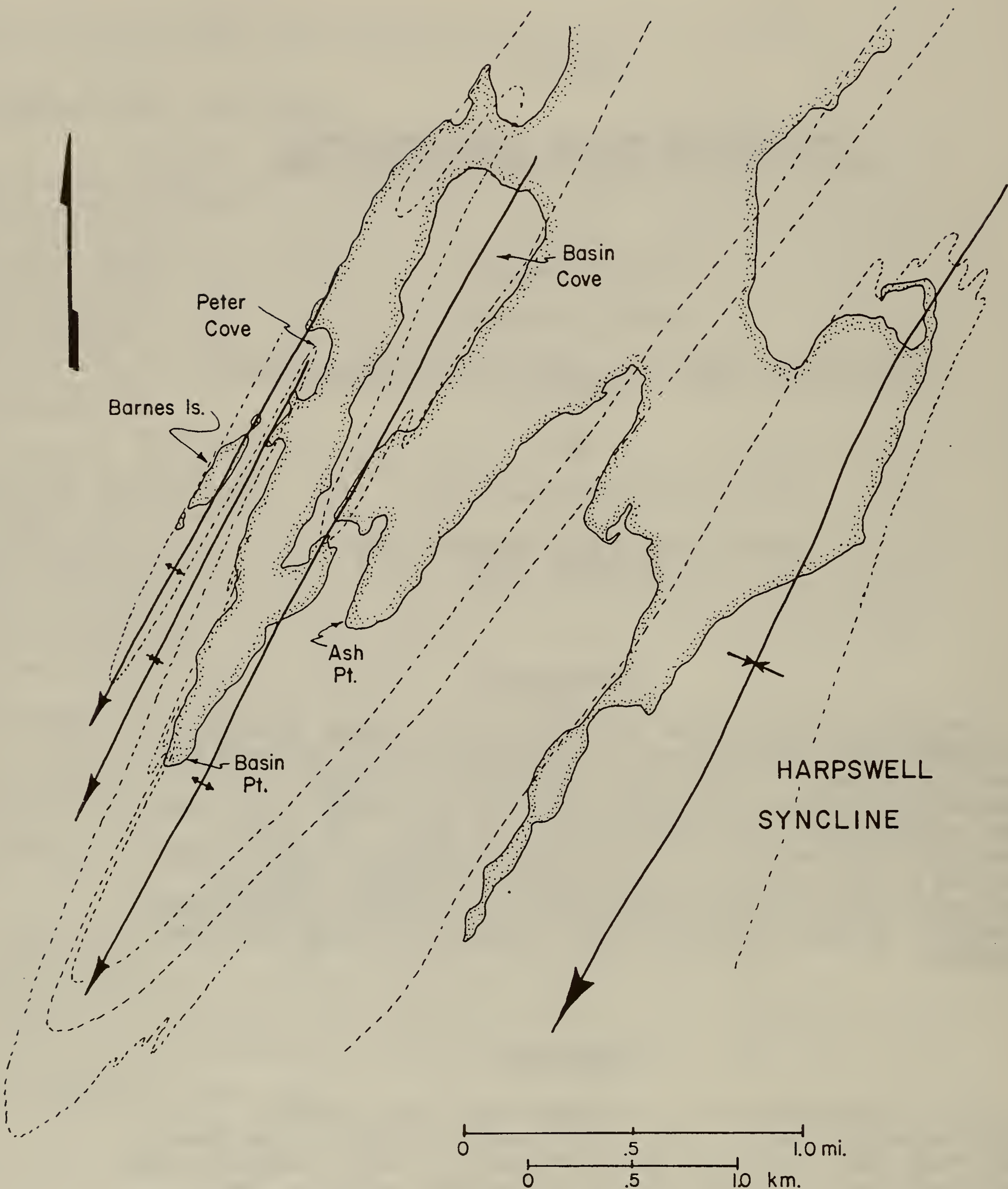

FIGURE 1: Regional structure of the Harpswell Neck area depicted as tight SW-plunging F2 folds. Field trip guide covers the NW shore of Harpswell Neck from Peter Cove to Basin Point and adjacent Barnes Island (after Hussey, 1971).

TABLE 1 - Summary of the stratigraphic relations and representative lithologies of the Casco Bay Group at Harpswell Neck

FORMATION	MAJOR LITHOLOGIES	MINOR LITHOLOGIES
JEWELL FORMATION	Muscovite biotite schist with abundant quartz vein boudins	
SPURWINK METALIMESTONE	Very fine grained medium dark gray, rusty weathering, biotite quartz pyrite chalcopyrite metalimestone	biotite quartzite calcareous biotite quartz schist calcsilicate quartzite
SCARBORO FORMATION	Fine grained medium dark gray to greenish gray muscovite biotite garnet quartz schist with thin discontinuous laminae and beds of muscovite biotite quartzite. Plagioclase quartz biotite magnetite schist.	amphibolite
DIAMOND ISLAND FORMATION	Very fine grained, dark gray to grayish black rusty, yellow and yellow-orange weathering, quartz graphite muscovite schist. Abundant thin (2mm) quartz veins parallel to schistosity.	
SPRING POINT FORMATION	Dark gray to grayish black, thin bedded, fine to coarse grained biotite garnet amphibolite Biotite garnet amphibole quartz plagioclase schist.	Fine grained, light gray quartz plagioclase biotite garnet amphibole gneiss and schist
CAPE ELIZABETH FORMATION	Very fine grained, light to dark gray, thin bedded muscovite biotite quartz feldspar +/- garnet schist. Very fine to fine grained light to dark gray, thin to medium bedded feldspathic muscovite biotite quartzite.	Calcsilicate gneiss and amphibolite.

modification and attenuation associated with the regional deformation.

The Casco Bay Group exposed at Harpswell Neck includes the Cape Elizabeth, Spring Point, Diamond Island, Scarboro, Spurwink and Jewell Formations. The units at Harpswell Neck exhibit very wide ranges in thicknesses, and in general, are much thinner than elsewhere in the Casco Bay area.

Jewell Formation - The Jewell is the most poorly exposed of the formations which crop out here. It is exposed at low tide within a small parasitic synclinal core between Barnes Island and the mainland. The Jewell, as observed here is a muscovite, biotite schist with abundant secondary quartz vein boudins.

Spurwink Metalimestone - The Spurwink is a fine-grained, medium dark gray quartz muscovite biotite metalimestone. Pyrite and chalcopyrite are locally important. Textures range from sugary granoblastic to schistose and gneissose. Layering is generally thin ($<$10 cm). Fish-mouth boudin structures are common. Overlying the metalimestone lithologies are thin (10 cm) to medium (30 cm) bedded quartzites with biotite and calcsilicate minerals.

Scarboro Formation - The Scarboro is a lithologically complex unit. Two lithologies are predominant. These are: 1) a poorly bedded or layered, very fine grained to fine grained muscovite, biotite garnet quartz schist with thin discontinuous laminae and beds of muscovite biotite quartzite and 2) a plagioclase quartz biotite magnetite muscovite schist. The latter occurs on the west side of Barnes Island stratigraphically above the former. In addition to these two major lithologies, thin beds of amphibolite are present. Stratigraphically, these are most common between the two dominant rock types observed here. Macroscopic textures of the Scarboro are distinctive. Thin discontinuous laminae of quartz and feldspar are common. These are locally observed as the noses or hinges of very small scale limbless folds. The origin of these is uncertain, but they may be interpreted to represent felsic blastopyroclasts within a meta-lithic tuff.

Diamond Island Formation - The Diamond Island is a fine to medium grained, dark gray to grayish black quartz graphite muscovite schist. The unit is characterized by rusty, yellow and yellow-orange weathering. Textures are uniform. Foliation is well developed. Thin ($<$ 2mm) discontinuous quartz lamellae are abundant and parallel regional schistosity. Bedding is not observed within the Diamond Island. This unit exhibits extreme variation in thickness here. At the type locality at Spring Point in Portland, it is 35 m thick. At Harpswell Neck thicknesses range from 1 m to an inferred stratigraphic thickness of approximately 62 m in a synclinal core at Peter Cove.

Spring Point Formation - At Harpswell Neck the Spring Point is easily recognized by its prominent, thin (2 - 15 cm), and compositionally distinctive bedding. The Spring Point exhibits a wide range of textures ranging from very fine to coarse grained. Rock types of individual beds range from amphibolites to schists and gneisses. Minerals and mineral proportions are variable, bed by bed. The amphibolite mineralogies may include, in varying proportions, anthophyllite, cummingtonite, hornblende, plagioclase, biotite, quartz and garnet. The schists and gneisses may include plagioclase, quartz, biotite, garnet, hornblende and muscovite.

Cape Elizabeth Formation - The Cape Elizabeth is the lowest formation in the Casco Bay sequence that will be observed today. The

Cape Elizabeth is predominantly thin bedded (2 - 30 cm). Bedding is variably developed. Outcrops may exhibit either poorly preserved or well preserved beds of alternating lithologies. Alternating lithologies include: 1) muscovite, biotite, plagioclase, quartz, +/- garnet schist and 2) feldspathic muscovite biotite quartzite. Additionally, the schistose beds contain chlorite porphyroblasts which cross cut the dominant S2 schistosity. Sizes of these porphyroblasts range from less than 1 x 3 mm to 3 x 8 mm. These appear to be randomly distributed on the S2 surface. However, careful observation may indicate that they are aligned within three separate, weakly developed lineations on the S2 surface. These three weak lineations are not always observable on the same schistosity surface. A third rock type observed within the Cape Elizabeth is granoblastic or gneissose calcsilicates. These occur as thin beds (< 30 cm) or boudins.

METAMORPHISM

The lithologies at Harpswell Neck have been subjected to two metamorphic events. The first is a regional prograde event of the low pressure facies series type. The second is a retrograde event. The first prograde event metamorphosed the lithologies at Harpswell Neck to the andalusite-staurolite zone. The retrograde event was to the chlorite zone. One of us (A.M.H., 1971) mapped the regional prograde isograds. The staurolite-andalusite isograd is to the southwest of the field trip area, while the staurolite-sillimanite isograd is to the northeast. Garnet is the only metamorphic index mineral from the first metamorphic event that has been observed, to date, in the exposures examined on this trip. Both staurolite and andalusite have been observed in appropriate lithologies elsewhere (Hussey, 1971). The second retrograde event produced chlorite porphyroblasts which cross cut the S1 schistosity. These are generally small, but porphyroblasts to 3 x 8 mm have been observed in schists of the Cape Elizabeth. The Cape Elizabeth Formation contains the best developed macroscopic evidence of this retrograde event. However, careful observation of the Spring Point and Scarboro Formations will disclose small chlorites which cross cut the S2 schistosity. This retrograde event may be related to the later phases of deformation. The chlorites have also grown across crenulations (F3) and their associated schistosity (S3). Undeformed chlorite porphyroblasts are also observed within sheared Cape Elizabeth lithologies at out-of-sequence contacts. This sheared, metamorphosed contact is between the Diamond Island and Cape Elizabeth lithologies along the NW shoreline of Harpswell Neck.

The age of the prograde metamorphism is interpreted to be Acadian (Osberg, Hussey and Boone, 1985). Thompson and Guidotti (1986) indicate that metamorphism of Carboniferous age affected much of southern Maine. It may be speculated that the retrograde metamorphism is related to the Carboniferous event. Conclusive documentation, is however, lacking at this time.

STRUCTURE

The overall structure of the Harpswell Neck area as illustrated

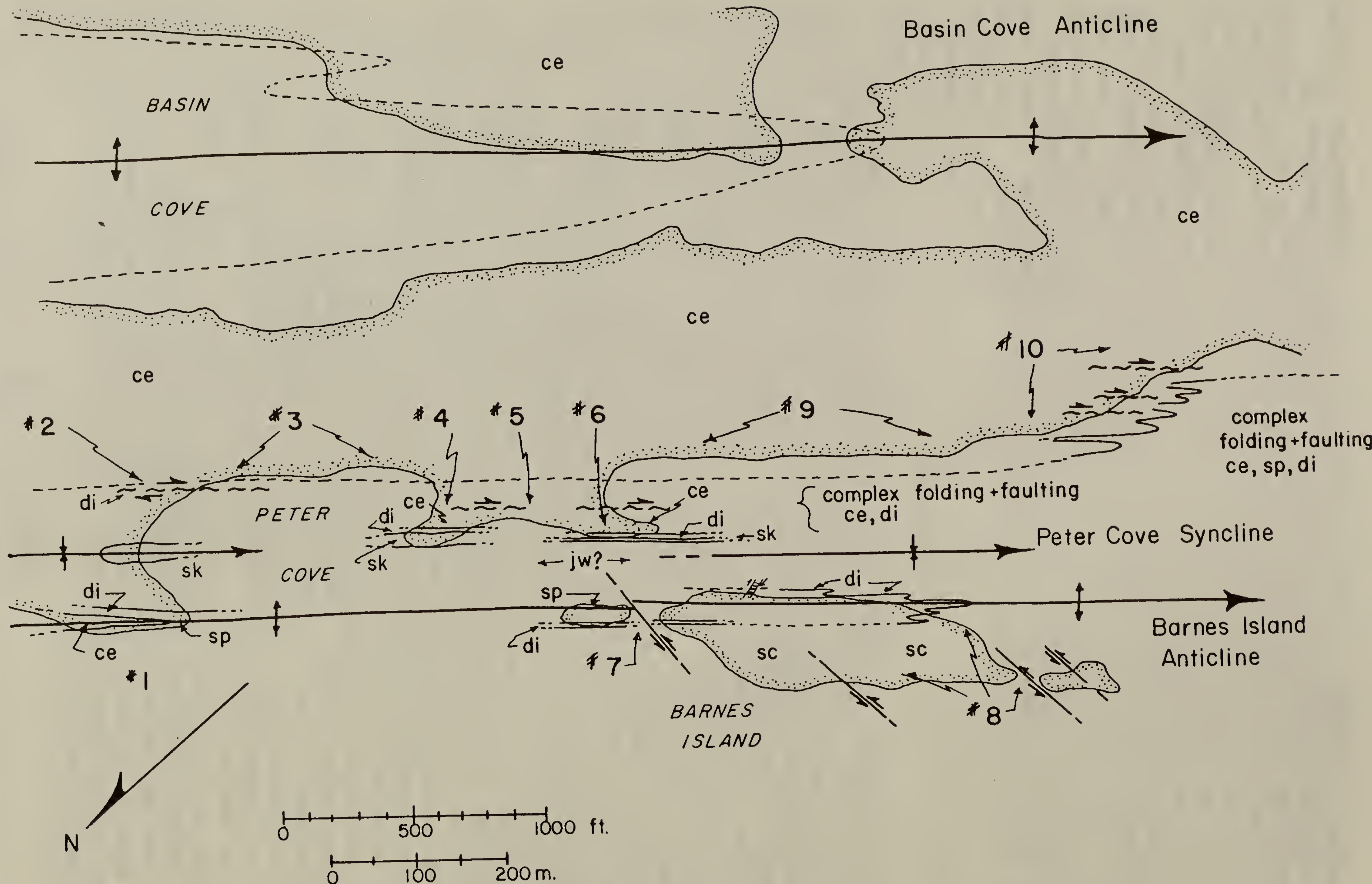

FIGURE 2: Geology and field trip localities for the NW coast of Harpswell Neck from Peter Cove to Barnes Island. All foliations and lithologic contacts are near-vertical and NE-trending.

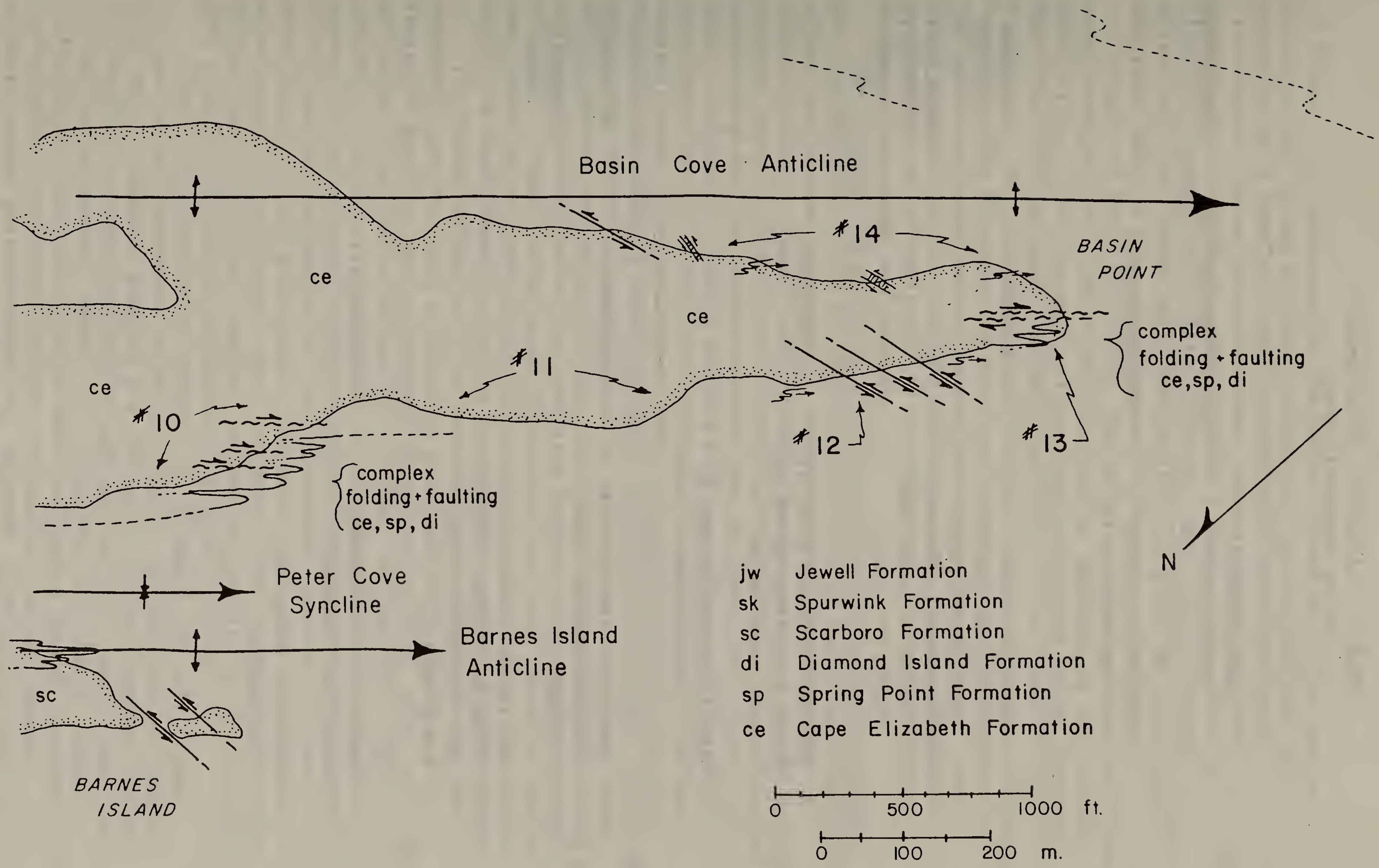

FIGURE 3: Geology and field trip localities for the NW coast of Harpswell Neck from Barnes Island to Basin Point. All foliations and lithologic contacts are near-vertical and NE-trending.

TABLE 2 - Structural Development in the Harpswell area

STRUCTURAL EVENT	CHARACTERISTIC STRUCTURES	AGE OF DEFORMATION	METAMORPHISM
F1	RECUMBENT FOLDS (not represented in study area)	PRECAMBRIAN(?)	(?)
F2	UPRIGHT TIGHT-ISOCLINAL FOLDS (gentle SW plunge) LAYER-PARALLEL DEXTRAL STRIKE SLIP ALONG LESS COMPETENT UNITS	ACADIAN	REGIONAL PROGRADE to ANDALUSITE STAUROLITE ZONE
F3	STEEPLY PLUNGING SCAR-LIP FOLDS & DEXTRAL ASYMMETRIC DRAG FOLDS MULTIPLE CRENULATIONS OF F2 SCHISTOSITY	ALLEGHANIAN	-
F4	CONJUGATE KINK SET WITH SINISTRAL STRIKE SLIP DOMINANT. RELATED TO LAYER PARALLEL DEXTRAL STRIKE SLIP IN BRECCIA FAULT STRUCTURES.	ALLEGHANIAN	RETROGRADE to CHLORITE ZONE
F5	KINK SET WITH NORMAL DIP SLIP DOMINANT. RELATED TO NW-SE EXTENSION.	MESOZOIC(?)	-

in Figure 1, is dominated by the Barnes Island anticline and the adjacent limb of the Harpswell syncline. These structures are separated by a zone of tightly folded upper Casco Bay Group lithologies. These rocks have been subjected to significant layer-parallel extension, dextral simple shear, attenuation of lithologic layers and brittle faulting. This area of intense extensional and simple shear deformation along the NW shoreline of Harpswell Neck is termed the Harpswell Neck high strain zone.

Figures 2 and 3 illustrate the detailed lithologic and structural relationships within the field trip area as outlined in the outcrop descriptions. The regional deformational structure of the Harpswell Neck area involves gently SW-plunging tight to isoclinal F2 fold structures. All schistosity surfaces are near vertical and northeast trending. The recumbent F1 fold structures recognized elsewhere in the Casco Bay Group are not recognized within the Harpswell area. Table 2 summarizes the interpreted sequence of structural phases involved in the deformation. Orientation data for these strutural phases is presented in Figure 4.

The regional SW-plunging isoclinal F2 fold structures (Fig.4a) are responsible for the outcrop pattern of Casco Bay Group lithologies within the Harpswell Neck area. Repetition of lithologic units across strike is common (see figure 5) in this area and can be related to tight parasitic F2 fold structures as well as superimposed dextral strike-slip faulting associated with the Harpswell Neck high strain zone. Lithologic contacts exposed within complex parasitic fold structures can be: 1) strictly conformable, gradational lithologic contacts, and 2) nonconformable out-of-sequence (fault) contacts. These out-of-sequence contacts, which juxtapose formations, are of two types. The first is distinguished by several centimeter wide zones which contain sheared and locally schistose lithologies. The second is distinguished by distinctive indurated breccia with complex internal structure. These contacts are interpreted as fault contacts, and in some cases as thin neck zones between mega-boudins where the boudins belong to those formations which exhibit more competent lithologies.

The Barnes Island anticline and the flank lithologies of the Harpswell syncline have developed a prominent SW-plunging b-axis lineation parallel to the regional F2 fold structures (Fig. 4c). This pervasive lineation is defined by a variety of features. These include: 1) mineral alignment (which includes biotite, anthophyllite, quartz, trains of small euhedral garnets and chert-like aggregates), 2) ribbing and corrugation of the schistosity surface, 3) pencil structure, 4) bedding-schistosity intersections, and 5) alignment of minor parasitic fold axes. Many of these lineations are interpreted to represent significant regional extension parallel to the F2 fold axes as is reflected in the prominent b-axis lineation.

Other structures which support our argument for significant regional b axis extension include a variety of boudinage structures (Fig. 4b). Fishmouth boudins are abundant. Features associated with the "mouths" of these boudins include open-pucker, tight-lipped and curled-lip structures. Quartz vein boudinage and planar, quartz filled parting surfaces occur between developing boudin structures and quartz veins. Partitions between the boudins contain euhedral garnets, coarse grained muscovite, and minor biotite, actinolite, quartz and calcite. These euhedral porphyroblasts as part of the

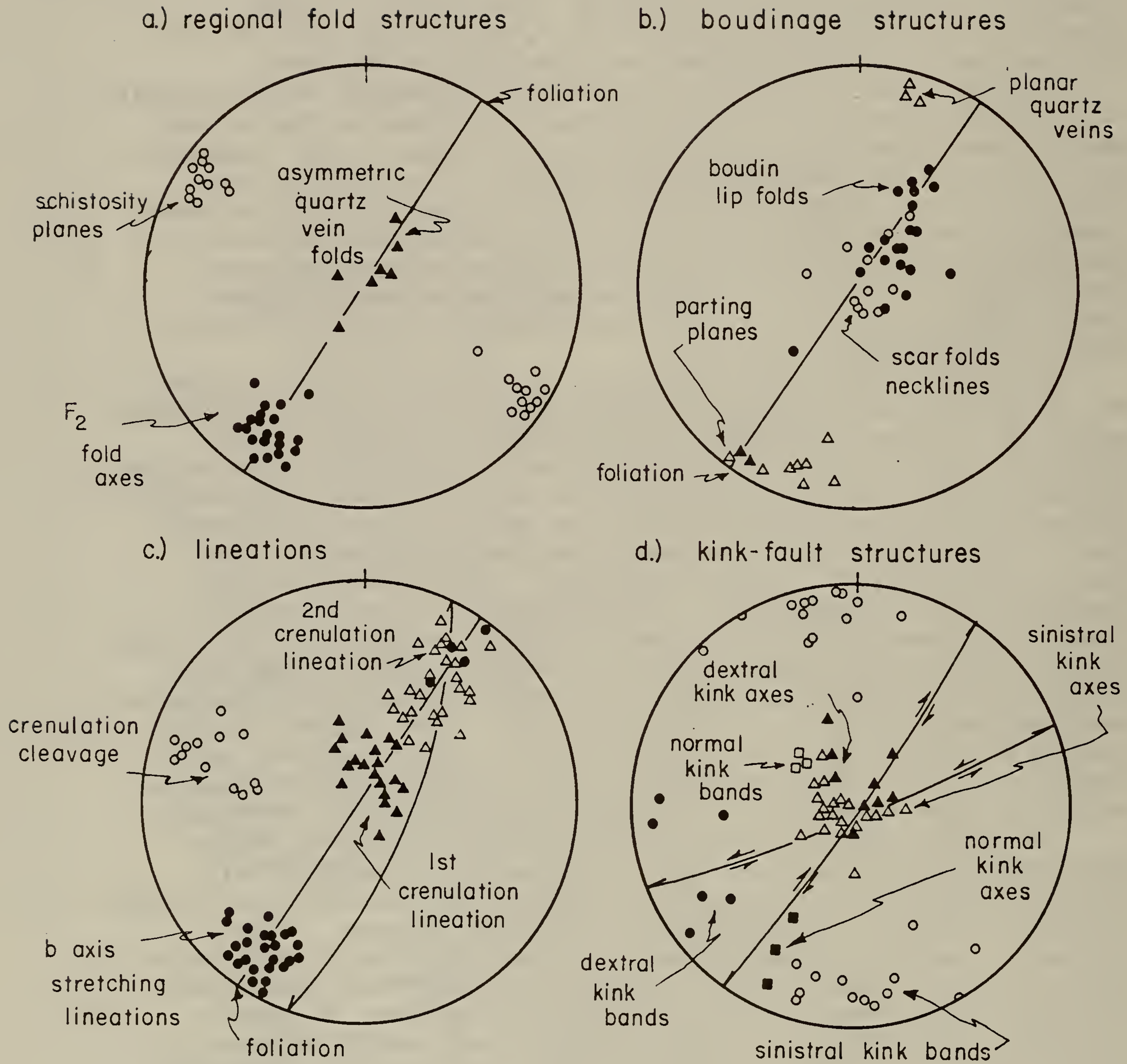

FIGURE 4: Stereographic projection (lower hemisphere) of structural data for the Harpswell Neck area. Structures plotted as planes (foliations, cleavage, faults), poles to planes (schistosity, crenulation cleavage, quartz veins, boudin parting planes, and kink bands) and lineations (fold axes, neck lines, kink axes and intersection lineations).

boudin partition mineralization suggests a temporal relationship between the regional prograde metamorphism and the development of the prominent b-axis lineations. The development of distinctly asymmetric foliation and quartz vein boudinage and segmented vein boudinage suggest strong layer-parallel dextral shear, a well as extension. This is particularly evident along the NW coast of Harpswell Neck through Peter Cove.

This zone of high extensional and dextral shear strain accommodation also contains both ductile (syn-metamorphic) and brittle faults. This assemblage of ductile and brittle structures represents a distinctive feature within this portion of Casco Bay, here termed the Harpswell Neck high strain zone. Brittle faults exhibit asymmetric shear zone fabrics and textures. Both types of faults document a long history of deformation. In addition to the faults, prominent dextral z-shaped quartz boudins, dextral offset quartz veins and structures within outcrop-scale brittle faults serve as the kinematic indicators which suggest that this was a broad zone of dextral shear strain.

The regional b-axis extension and layer-parallel ductile dextral shear produce steeply-plunging asymmetric folds (F3) and associated structures that clearly deform the F2 axial surfaces. These folds consist of lip folds and scar folds associated with the boudinage and asymmetric folded quartz veins and boudins as well as z-shaped asymmetric intrafolial folds many of which contain an oblique axial plane crenulation cleavage expressed as lineations on the schistosity surfaces. Exposures exhibit several possible sets of crenulations. These are expressed as multiple lineations on the S2 schistosity surfaces of the more micaceous units, particularly within the Cape Elizabeth Formation (Fig. 4c).

The F4 fold structures consist of brittle steeply-plunging kink folds and kink bands, of both dextral and sinistral sense, which have a direct spatial association with numerous EW-trending sinistral strike-slip faults and layer-parallel dextral strike slip faults that cut both the Barnes Island and Harpswell Neck exposures. These faults and kinks are interpreted to be a coherent structural assemblage associated with the later deformation history of the Harpswell Neck high strain zone.

The F5 structures include an additional set of kink bands with horizontal fold axes (Fig. 4d). These kink structures are interpreted to express a NW-SE extension most likely associated with Mesozoic tectonic events represented within the study area by a single basalt dike exposed along the NW shoreline of Harpswell Neck and minor NE-trending normal fault structures.

REGIONAL CORRELATIONS

The Barnes Island anticline illustrates the clear stratigraphic succession of the Cape Elizabeth, Spring Point, Diamond Island and Scarboro Formations within the Casco Bay Group. Significant shear and/or extensional strain has been superimposed on all lithologies as represented by asymmetric structures and fabrics and the prominent b-axis elongation lineation. There is a close spatial and temporal association between the SW-plunging regional F2 fold structures and the regional extension. These structures may be related to the

deformational history of the Casco Bay formations within the regional Norumbega fault system as a series of originally en echelon fold structures. These F2 folds have been tightened to upright isoclinal structures and sheathed by regional extension and dextral shear strain associated with the Norumbega fault system.

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ITINERARY

ASSEMBLY POINT: Basin Point at Dolphin Marina, Harpswell Neck. From Brunswick, take Route 123 (Harpswell Road) south for approximately 12 miles. Turn right onto Ash Point Road at signs for Dolphin marina and an elementary school. Take next right onto Basin Point Road following signs to Dolphin Marina. Enter marina and proceed through boatyard to grassy area at Basin Point. Time : 8:00 A.M. The group will then proceed to the first outcrop. Access to the exposures is dependant on the tidal cycle. Because of this it may be necessary to visit the exposures in a sequence different from that listed below.

See Figures 2 and 3 for outcrop location. All schistosities are NE trending and near vertical within the field trip area.

LOCALITY 1: Anticlinal core axis at northwestern end of Peter Cove.

This exposure represents the deepest section of the gently SW-plunging Barnes Island anticline. Formations present here include

Cape Elizabeth, Spring Point and Diamond Island. The Cape Elizabeth consists of thin to medium beds of quartzite or quartz mica schist. These quartzites are exposed within the northernmost hinge zone of the Barnes Island anticline. Conformably overlying the Cape Elizabeth Formation is the Spring Point Formation. Spring Point lithologies include well bedded very fine to coarse grained, amphibolites, schists and gneisses. The more common lithologies include anthophyllite garnet plagioclase quartz amphibolites, biotite garnet plagioclase quartz schists, and plagioclase quartz biotite schists. Minor asymmetric folds, within the Spring Point, can be seen on both limbs of this well-defined SW-plunging anticlinal structure. Trains of euhedral garnets are locally observed to produce both "b" and "a" lineations. A conformable contact between the Spring Point formation and the overlying Diamond Island Formation is exposed on the SE limb of the anticline adjacent to Peters Cove. The Diamond Island is also exposed as a 30 cm wide layer within the core of a small parasitic syncline, also on the southeastern limb of the anticline.

Approximately 125 m of the Diamond Island Formation is linferred to be present under the beach at Peter Cove. A small outcrop of silicified metalimestone crops out in the middle of the beach area. This represents the Spurwink Metalimestone. The presence of Spurwink, combined with the unusual thickness of the Diamond Island suggest to us that the Peter Cove beach area, and Peter Cove is underlain by the thickened nose of a SW-plunging F2 syncline.

LOCALITY #2: Peter Cove fault zone-

Two types of faults are present here. One is interpreted as a pre or syn-metamorphic fault, the other is a post retrograde brittle fault.

The first fault produces an out-of-sequence contact between Diamond Island and Cape Elizabeth lithologies. Note that the Spring Point Formation which was well exposed on the opposite side of the beach is completely missing here. The contact between the Diamond Island and Cape Elizabeth is sharp and parallels the regional schistosity. There is no apparent gouge or breccia. Later today we will have the opportunity to examine this same fault where 20 cm thick boudins of Spring Point separate the Diamond Island from the Cape Elizabeth.

The post retrograde metamorphic fault is exposed as a meter-wide zone of indurated fault breccia. The breccia consists of Diamond Island lithologies. This fault zone is located near the base of the cliff at the northeastern end of Peter Cove. This NE-trending fault breccia zone is developed parallel to the foliation in the adjacent Cape Elizabeth and Diamond Island lithologies. Additional exposures of the same fault zone will be examined along strike to the SW. This fault represents a late brittle layer-parallel dextral strike-slip fault of limited displacement. This fault is within 2 m of the previously faulted out-of-sequence contact between the Cape Elizabeth and Diamond Island Formations. Kinematic indicators in this brittle fault exposure include subhorizontal slip lineations along breccia fragment surfaces, internal minor dextral strike slip surfaces, and small-scale drag folds and high-angle kink structures.

LOCALITY #3: Cliff exposures along shore of Peter Cove-

High-standing cliffs consisting of muscovite quartz schists, muscovite quartzites and calcsilicates of the Cape Elizabeth Formation are exposed along this traverse. The Cape Elizabeth exhibits a well-developed SW-plunging b-axis elliptical or ovoid lineation on the near-vertical axial plane (F2) schistosity. These chert-like ellipses or ovoids consist of very fine grained (cryptocrystalline) material which appears opaque in thin section.

Planar quartz veins, oblique-boudined quartz veins and puckered fishmouth foliation boudinage with accompanying asymmetric scar folding can be observed locally. Steeply-plunging crenulation lineations within the near-vertical schistosity are well developed as intersection lineations of an oblique crenulation cleavage associated with moderately NE-plunging z-shaped minor F3 folds.

Along this traverse the faulted contact between the Cape Elizabeth and Diamond Island Formations is again exposed. This interpreted fault structure consists of a broad zone of sheared lithologies and abundant boudined quartz veining. Sheared Cape Elizabeth lithologies contain fine-grained, blackish, chert-like, elliptical to rod-shaped features, as well as flattened garnet porphyroblasts. Spring Point lithologies may be preserved as minor remnants within the sheared contact zone.

LOCALITY #4: Silicified marble with sulfide mineralization-

This unusual lithology is interpreted to represent the Spurwink Metalimestone. Lithologies consist of a rusty-weathering silicified impure marble with abundant sulfide (pyrite and chalcopyrite) mineralization. The marble exhibits well-developed pucker and open-fishmouth foliation boudinage. The prominent sulfide mineralization occurs as a partition mineralization along with quartz and coarse grained muscovite. The sulfides also occur as cross-cutting veins and as a disseminated intergranular components within some horizons. This lithologic unit is flanked on both sides by the Diamond Island Formation. These out-of-sequence contacts between the Spurwink and the Diamond Island (Scarboro is missing) and between the Diamond Island and Cape Elizabeth formations (Spring Point missing) can be explained by faulting and/or large-scale boudinage to account for the missing Scarboro and Spring Point lithologies.

LOCALITY #5: Black graphitic and muscovite quartz garnet schists-

This series of outcrops consists of well-bedded muscovite quartz feldspar garnet schists of the Cape Elizabeth Formation and blackish quartz graphite muscovite schists of the Diamond Island Formation. These lithologies are interpreted as lithotectonic units in out-of-sequence formational contact. These contacts show significant layer-parallel extension and dextral shear which is represented by symmetric and asymmetric foliation and quartz vein boudinage. Some boudinage of the quartz vein material has produced flow folding (scar folding) of the schistose host into the separation zone between boudin segments. Several scar folds are Z-shaped and locally develop their own axial plane cleavage. Several steeply-plunging asymmetric Z-shaped quartz vein folds are exposed. These support a dextral,

non-coaxial simple shear strain history for the Harpswell Neck high strain zone.

Contacts between the Cape Elizabeth and Diamond Island lithologies are interpreted as faults. These faults include abundant quartz veining, boudinage and distortion of schistosity surfaces. Younger brittle strain is localized along these contacts. These younger brittle fault breccia zones are represented by the along strike continuation of the fault breccia zone in locality #2. Subhorizontal slip lineations along the breccia fragment surfaces supports a strike-slip (dextral) interpretation for this fault structure.

LOCALITY #6: End of access road to Barnes Island-

Lithologies include the blackish quartz graphite muscovite schist of the Diamond Island Formation, muscovite quartz garnet schists and micaceous quartzites of the Cape Elizabeth Formation, and a thin ductilly deformed metalimestone remnant of the Spurwink. The Jewell Formation may be present in poorly exposed low tide outcrops just offshore toward Barnes Island.

These exposures contain abundant stringers of asymmetric quartz vein boudins which generally transgress through the schistosity, and compositional layering at a slight oblique angle. This relationship would be expected for the reorientation and extension of discordant quartz veins during subhorizontal layer-parallel dextral simple shear. Distinctly asymmetric foliation boudinage within the Spurwink is also indicative of dextral strike-slip shear strain. The texture of the Spurwink Metalimestone can also be related to grain size reduction processes during ductile flow or mylonitization. A considerable amount of dextral simple shear strain has been accommodated within these lithotectonic units.

A SW-plunging mineral smear lineation is prominently developed parallel to the regional fold axes. Scar folding of the host lithologies about the boudinage is common. A single mullion structure is also developed in these exposures.

Additional layer-parallel fault structures are evident. These bring the Diamond Island into contact with the Cape Elizabeth. One of these contacts is expressed as a foliated boudined fault zone. This fault appears to contain lithologic remnants suggestive of the Spring Point Formation. This out-of-sequence contact may be the result of large-scale boudinage of the more competent Spring Point lithologies. The Spurwink itself may represent an early ductile fault or thin boudined neck which removes the Scarboro and juxtaposes the Spurwink with the Diamond Island formation.

Exposures along the backside of this small peninsula at the adjacent cove beach represent the continuation of the brittle fault breccia zone seen in the previous locations. Dextral shear strain is localized along the margins of this meter-wide limited-displacement fault zone with little internal structural development or brecciation. A conjugate set of steeply-plunging F4 kink band structures is also exposed here.

Barnes Island exposures-

LOCALITY #7: Saddle exposures and small island at NE end of Barnes Island-

The Spring Point is the dominant formation in these exposures. Rock types consist of thin beds of biotite garnet plagioclase quartz schists, anthophyllite garnet plagioclase quartz amphibolite and quartz biotite garnet schist. Locally, the fibrous anthophyllite is aligned parallel to the prominent b-axis lineation. This is interpreted to represent grain growth during regional extension. Several exposures (NE end of Barnes Island) contain recognizable bedding in parasitic folds with an accompanying axial plane schistosity. These are easily observed on the SE limb of this SW-plunging anticline.

The axis of the main Barnes Island anticline trends along the SE side of Barnes Island. The Diamond Island is approximately 1 m thick on the west limb of the anticline, and substantially thicker on the east limb. The Scarboro Formation is in excess of 50 m thick, and crops out in the center and west side of the island. These exposures constitue the NW limb of the Barnes Island anticline.

Exposures in the saddle area between the two islands contain numerous EW-trending sinistral kink structures up to a meter in width. These sinistral kinks occur in association with discrete sinistral strike slip faults. Displacements along these faults are estimated to be less than 10 m. Smaller-scale dextral kinks are also present in conjunction with minor layer-parallel dextral strike-slip surfaces. A single 3 meter wide dextral kink structure is exposed to the south of the saddle area on the eastern side of the island. These structures may be correlated with the layer-parallel dextral strike-slip faults and fault breccia zones seen in previous outcrops. An additional layer-parallel dextral strike-slip fault is exposed along the SE shoreline within the Diamond Island and Spring Point Formations.

LOCALITY #8: SW end of Barnes Island-

Lithologies in these exposures consist dominantly of thin and poorly bedded and laminated muscovite biotite quartz garnet schist and minor amphibolite of the Scarboro Formation. Tight parasitic asymmetric folds with gentle SW plunge are well developed in these exposures. The repetition of Scarboro, Diamond Island and Spring Point lithologies along the very southern end of Barnes Island is interpreted as a series of tight isoclinal parasitic fold structures. All plunges are to the southwest.

Considerable alteration and actinolite mineralization is developed within the Scarboro Formation as part of the regional b-axis extension and boudinage. Partition mineralization associated with the boudinage consists of quartz, coarse grained muscovite and biotite, chlorite, actinolite and minor euhedral coarse grained garnet. A prominent b-axis extension is expressed as strongly lineated quartz sheets that define the schistosity within the Scarboro Formation. Flattened and distinctly asymmetric magnetite porphyroblasts are developed within plagioclase quartz biotite magnetite muscovite schists of the Scarboro along the NW shoreline of Barnes Island. These flattened porphyroblasts are evidence for dextral strike-slip

shear strain associated with the prominent b-axis lineation.

These exposures of the Scarboro Formation particularly along the NW shoreline contain the best developed fishmouth boudinage structures in the area. These foliation boudin structures vary from mere puckers to tight-lipped closed-mouthed pseudo-fold structures and to dramatic curled-lip fishmouth structures with steeply-plunging lip folds.

Saddle exposures between the main portion of Barnes Island and the smaller SW end of the island contain sinistral strike-slip faults and sinistral kinks similar to those observed in previous exposures. These brittle structures have controlled the morphology of the island in creating the cross-cutting topographic saddles at either end of the main island.

Harpswell Neck Exposures

LOCALITY # 9: High cliff face exposures along the NW Harpswell Neck shoreline-

The high standing cliff exposures along the NW shoreline of Harpswell Neck consist dominantly of muscovite quartz biotite garnet schist, micaceous quartzite and minor calcsilicate of the Cape Elizabeth Formation. The axial plane schistosity is well developed here and contains a prominent b-axis lineation expressed as extremely elongated fine-grained chert-like inclusions up to 30 centimeters in length. The steeply NE-plunging crenulation lineations are absent while a gently NE-plunging crenulation lineation is locally present in association with Z-shaped asymmetric minor F3 folds. A large erratic boulder of syenite is present along the shore. A single 20 cm. wide Mesozoic basalt dike has also intruded parallel to the prominent schistosity.

LOCALITY #10: Repeated lithologic units and fault slice exposures at change in trend of shoreline-

As the trend of the shoreline changes toward the SW, exposures consist of several distinctive lithologies which include representatives of the Diamond Island, Spring Point and Cape Elizabeth Formations. These lithologies are repeated across strike by minor parasitic isoclinal folding and/or layer-parallel strike-slip faulting. Both conformable gradational contacts and sheared out-of-sequence contacts can be observed in these exposures. The conformable contacts and observed minor asymmetric folds help to recognize this section as a series of SW-plunging parasitic folds on the west flank of the Harpswell syncline (see Fig. 5).

Asymmetric sigmoidal quartz boudins are preserved in these exposures. These suggest a ductile phase of dextral simple shear and/or strike-slip faulting. Several larger boudins exhibit well-developed steeply-plunging asymmetric scar folds and steeply-plunging boudin neck lines. The northwesternmost exposure of the Spring Point may represent a large-scale boudin where out-of-sequence contacts have developed between the Cape Elizabeth and

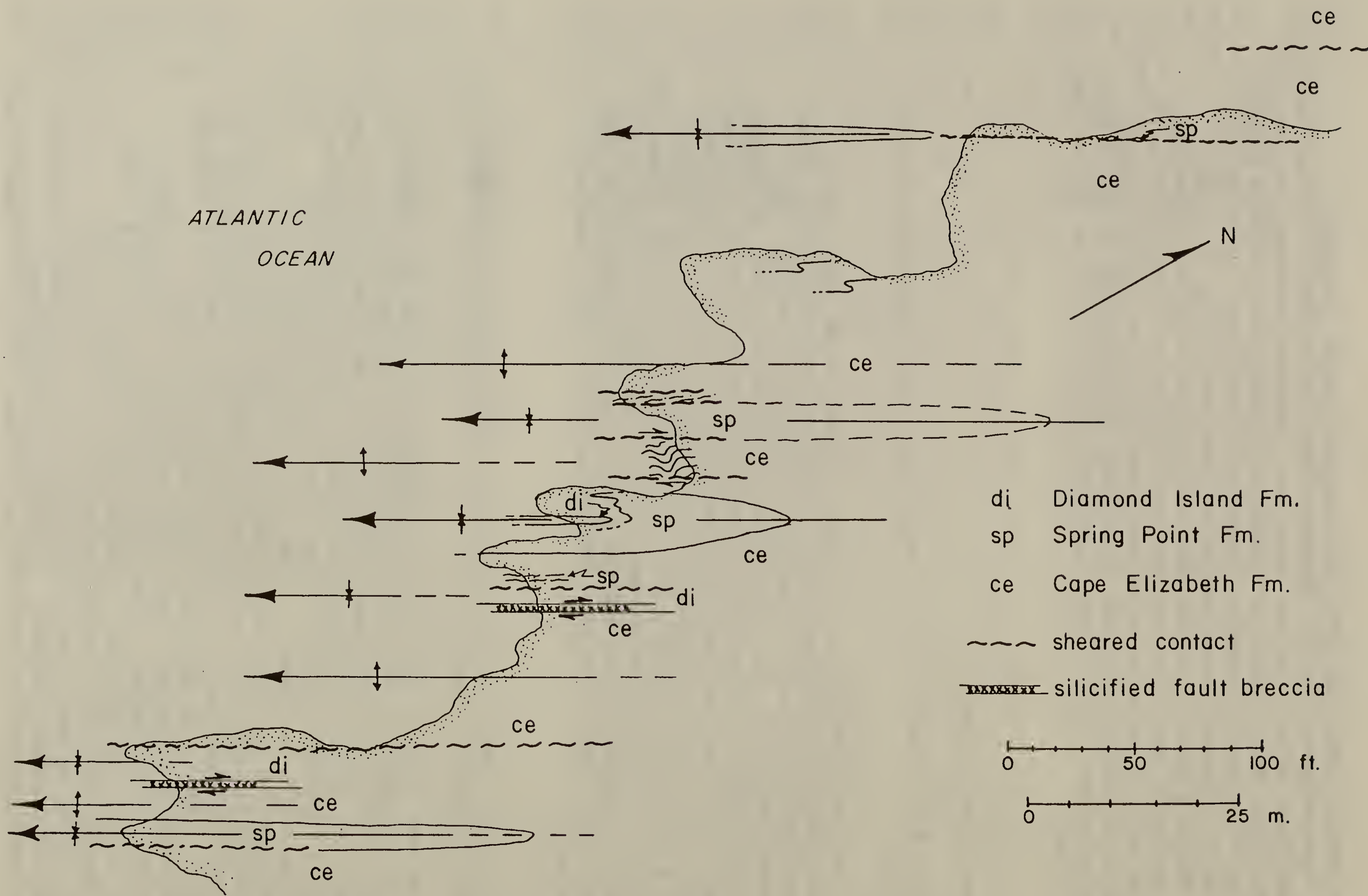


FIGURE 5: Detailed geologic map of a portion of the NW shoreline of Harpswell Neck between Peter Cove and Basin Point (Locality #10). All foliations and lithologic contacts are near-vertical and NE-trending.

the Diamond Island Formations within the neck areas of the boudins. Minor fibrous anthophyllite veins are preserved along with a distinct fibrous pull-apart texture. The aligned fibrous mineral growth and distinct dilational features are representative of the prominent b-axis extension. This extension is parallel to the regional SW-plunging F2 fold axes.

Schistose Cape Elizabeth lithologies contain abundant chlorite porphyroblasts up to 4 mm in length randomly oriented with respect to the regional F2 schistosity. Sheared Cape Elizabeth lithologies at out-of-sequence contacts with the Diamond Island Formation contain undeformed chlorite porphyroblasts. This indicates that layer-parallel faulting and/or large-scale boudin necking occurred prior to the retrograde chlorite metamorphic event.

Several out-of-sequence contacts in these exposures are marked by later layer-parallel breccia fault zones with indurated breccia, silicification, minor gouge development and complex, generally steeply-plunging internal fold structures. Quartz-filled en echelon feather fractures and steeply-plunging internal fold structures indicate dextral strike-slip faulting. One 2 meter-wide layer-parallel fault structure within the Cape Elizabeth lithologies contains steeply-plunging box fold structures similar to structures found at Basin Point.

Small centimeter-wide sinistral kinks with steeply-plunging rotation axes are developed in these exposures. A second set of kink structures with horizontal rotation axes are also represented. Minor normal faults of minimal displacement are present. These exhibit down-dip striations.

LOCALITY #11: Crenulations along the Harpswell Neck shoreline

The continuing coastal exposures along the NW Harpswell Neck shoreline consist exclusively of the Cape Elizabeth Formation. The exposed lithologies include thick-bedded quartzites, thin and poorly bedded, fine to very fine grained muscovite biotite quartz garnet schist and thin bedded calcsilicates. The quartzites exhibit well-developed S-shaped, SW-plunging F2 parasitic folds. The micaceous schists exhibit abundant boudined quartz veins and well developed crenulation lineations. These crenulation lineations occur in two sets. The dominant set is a steeply NE-plunging set that is cross-cut by the second set which plunges gently to the NE. The two sets of crenulations have developed prior to the retrograde chlorite metamorphic event. Chlorite porphyroblasts have grown across the crenulations. The crenulation lineations represent an intersection lineation that is caused by the intersection of an oblique crenulation cleavage and the F2 schistosity. The crenulation cleavage is developed axial planar to open NE-plunging Z-shaped minor F3 fold structures.

LOCALITY #12: Brittle fault exposures along NW Harpswell Neck shoreline-

The cliff exposures near Basin Point contain several well

developed ENE-trending brittle sinistral strike-slip faults. These cut the Cape Elizabeth Formation. These faults appear to be preceeded by a phase of F4 sinistral kinking. Continued slip along the kink boundaries developed into discrete strike-slip faults. These faults display well developed steeply plunging drag folds, subhorizontal striations and polished surfaces. These fault zones have also been the site for gouge formation, and silicification resulting in a prominent rusty weathering appearance. Displacements along these faults are probably on the order of several meters. The cumulative displacements of these faults may substantially effect the outcrop pattern within the Harpswell Neck exposures.

Boudin structures are particularly well-developed in these exposures. Quartz vein boudinage has developed clear steeply-plunging neck lines and associated scar folds. Subhorizontal extension of some of the thicker quartzite beds relative to the more micaceous layers within the Cape Elizabeth Formation has developed rectilinear quartz partition mineralizations. Continued subhorizontal extension with an apparent ductility contrast between the more competent partition mineralization and the quartzite layers has resulted in necking of the quartzite layers about the partition. Several of these rectilinear quartz partitions appear to be arranged en echelon and can be interpreted to be segmented earlier quartz veins. These veins, up to 50 cm. in width, are disrupted by dextral layer-parallel shear. Dextral slip has been concentrated along the more ductile schistose layers producing the offset quartz pod patterns observed in these exposures.

A single large silicified boudin pod approximately 2 meters in width has also been isolated by this boudin-producing deformation. This silicified boudin pod has been necked abruptly. Similar lithologies are not present along strike in these exposures suggesting significant regional extension.

LOCALITY # 13: SW tip of Harpswell Neck at Basin Point-

At the SW tip of Basin Point a sharp axial crest of a minor anticlinal F2 fold appears to be gently folded about the steeply plunging axis of boudin related scar folds similar to those seen elsewhere on this trip.

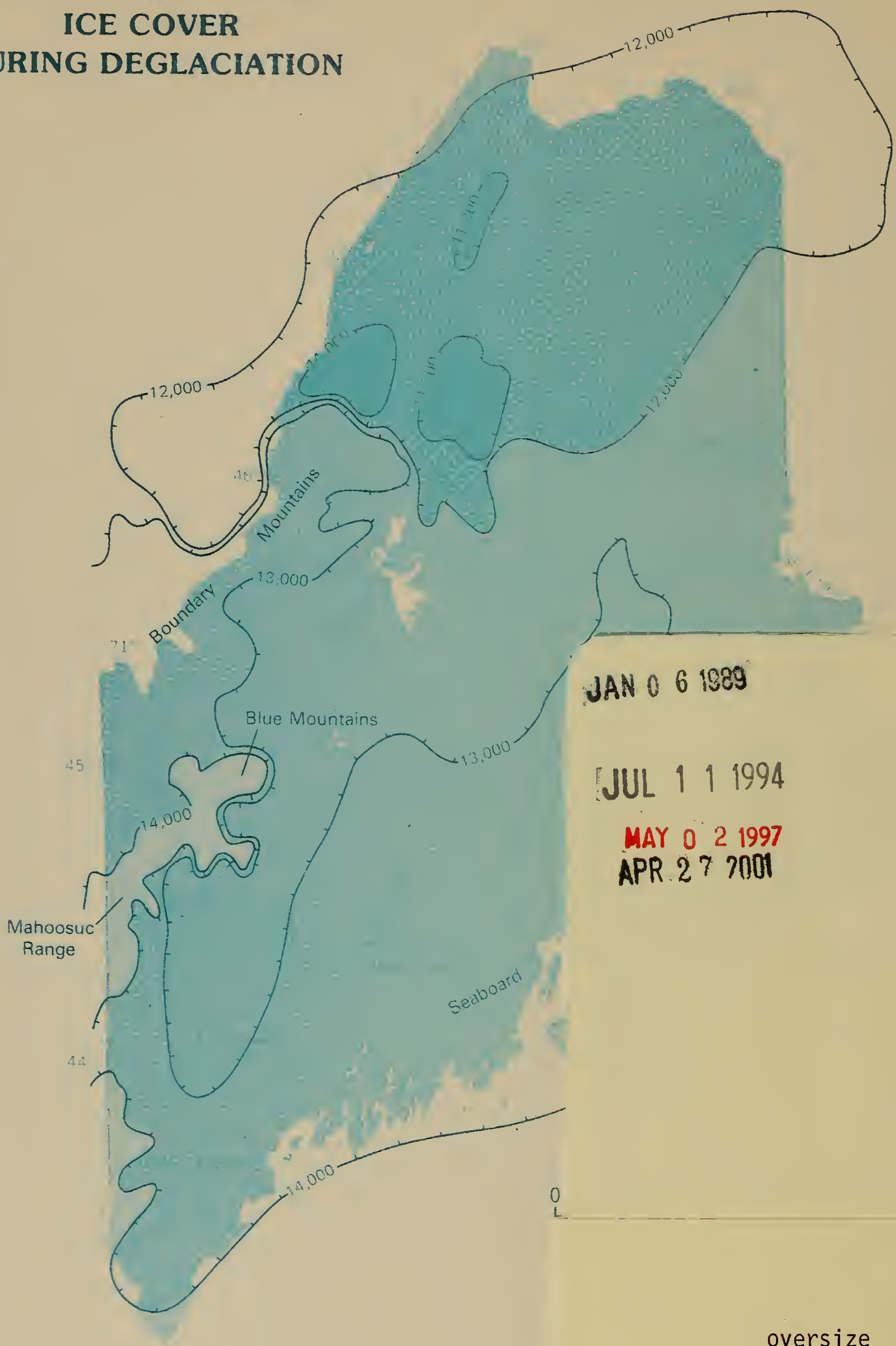
At Basin Point, several lithologies are present. These represent the Cape Elizabeth, Spring Point and Diamond Island Formations. Several conformable contacts between the Cape Elizabeth and Spring Point lithologies are present within tight synclinal F2 folds. These SW-plunging synclines expose the younger Spring Point within the cores. The Diamond Island Formation is in apparent fault contact with the Cape Elizabeth. One of these faulted contacts is a 5 meter wide layer-parallel zone of contorted rock within the Cape Elizabeth Formation. The internal contortion involves steeply-plunging box and kink folds similar to those within fault zones at locality #10. This zone of faulted and parasitically folded lithologies may be correlated with similar structures and lithologies found farther to the NE along Harpswell Neck at locality #10 where they are now offset by the prominent sinistral strike-slip faulting described in locality #12.

LOCALITY #14: SE shoreline of Harpswell Neck-

The SE shoreline of Harpswell Neck consists exclusively of the Cape Elizabeth Formation. Lithologies are similar to those seen elsewhere in the Cape Elizabeth. These exposures exhibit well developed thin to thick beds. These lithologies exhibit well developed scar folding, quartz vein boudinage, and isolated steeply plunging quartz vein fold noses, often asymmetric and z-shaped. Large scale pencil structures are well developed here parallel to the regional SW-plunging F2 fold axes often a meter or more in length. Crenulation lineations are prominently developed in the thinner bedded units with the gentle NE-plunging set dominating. Garnet grains a observed to exhibit tails or pressure shadows of quartz. These tails or pressure shadows are commonly sigmoidal in shape suggesting crenulation or rotatation.

Dextral and sinistral centimeter-scale kinks are locally very prominent. These create distinctive chevron folds whose axes plunge subparallel to crenulation lineations. Larger-scale meter-sized kinks, planar quartz veins and sinistral strike-slip faults (one with over two meters of displacement) are also present in these exposures.

INFERRED EXTENT OF ICE COVER DURING DEGLACIATION



Map shows successive positions of the late Wisconsina
11,000 years ago (adapted from Davis and Jacobsor
Extent of ice cover is based on limiting radiocarbon da
topography. Number and configuration of ice remna
Only the outer limits of essentially continuous ice co
tions regarding ice dynamics. Some mountainous are
thinning ice sheet are omitted. Darker colors show p

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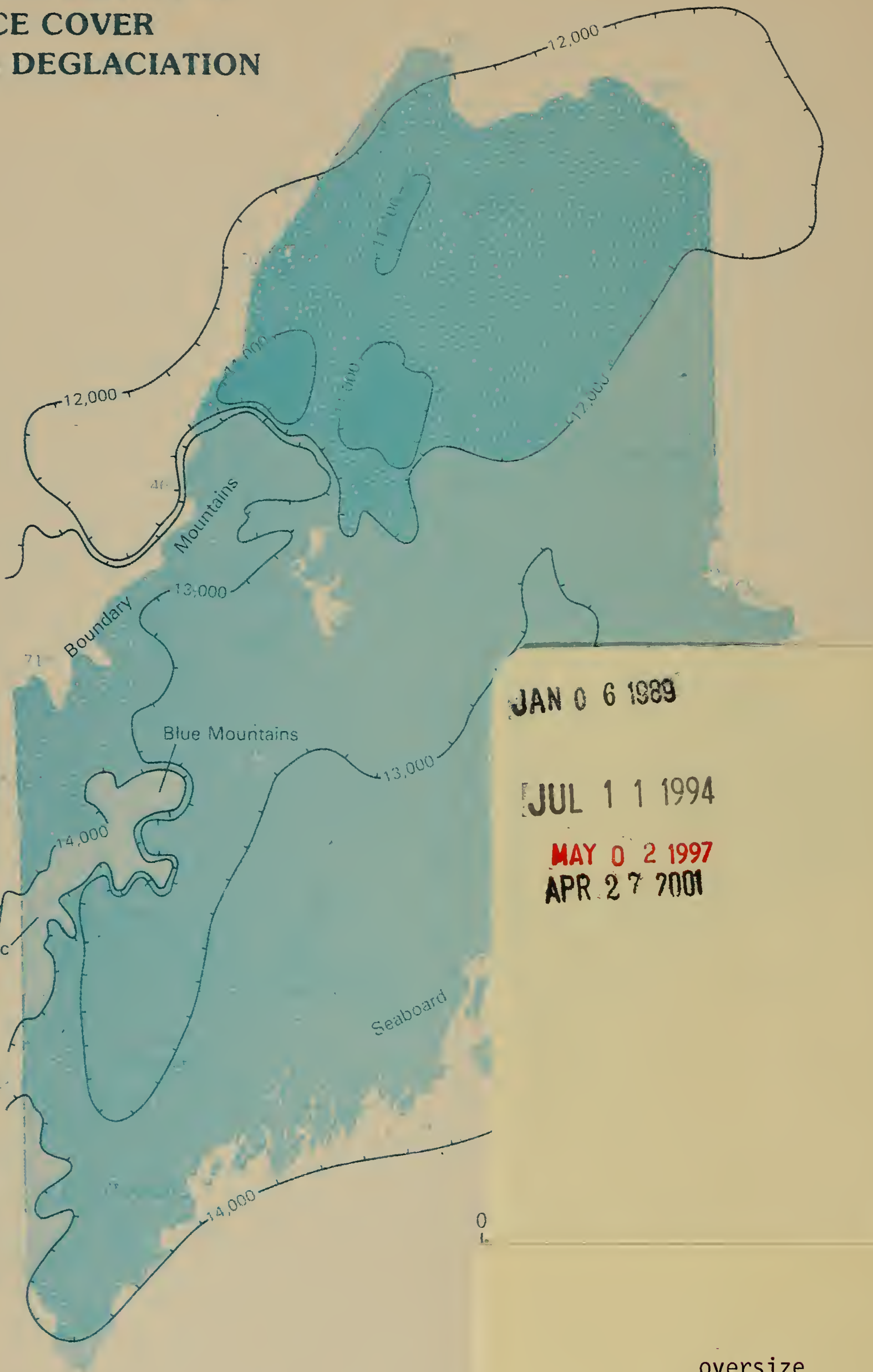

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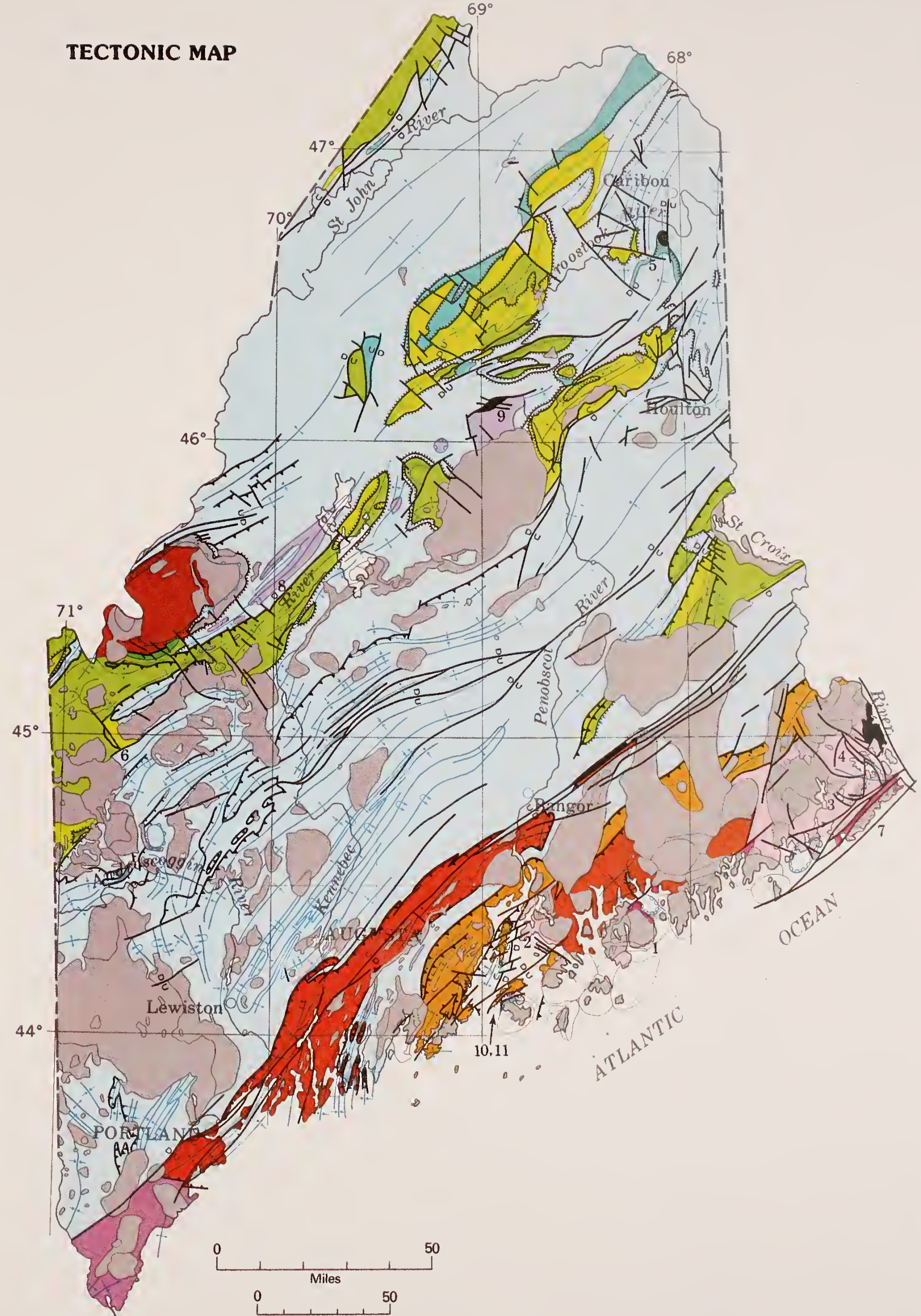
TECTONIC MAP



Post-orogenic Carboniferous and Devonian sedimentary rocks

Central and Western Maine

- Undivided Middle Ordovician to Lower Devonian metasedimentary rocks
- Middle Devonian (Emsian) volcanic rocks of the Piscataquis Volcanic Belt
- Silurian to Lower Devonian basaltic and andesitic volcanic rocks in north-central Maine
- Ordovician mafic and minor felsic volcanic rocks
- Cambrian to Lower Ordovician belts of mafic to felsic volcanic rocks and associated sedimentary rocks; mélange prominent north and west of the Kearsarge-Central Maine Synclinorium
- Ophiolitic rocks of Cambrian age
- Cambrian Grand Pitch Formation of the Weeksboro-Lunksoos Lake anticlinorium
- Precambrian Z to Silurian rocks of the Merrimack Trough
- Precambrian Chain Lakes terrane

Coastal Lithotectonic Block

- Silurian to Lower Devonian metasedimentary rocks
- Silurian to Lower Devonian mafic to felsic volcanic rocks
- Precambrian Z to Ordovician metasedimentary and metavolcanic rocks of the Cookson-Penobscot belt (includes North Haven-Islesboro sequence)
- Precambrian Z to Ordovician metasedimentary and metavolcanic rocks of the Casco Bay Group and Ellsworth and Columbia Falls Formations

The descriptions of the tectonic units are not necessarily the dominant lithology in an area, but are lithologies significant for tectonic interpretations. The dominant lithologies are shown on the geologic map.

5 Volcanic formation with isotopic age determination; refer to list below for reference(s).

Volcanic Formations with Isotopic Age Determinations

1 Bar Harbor Formation (46,157)
2 Castine Formation (45,46)
3 Dennys Formation (31,76,77)
4 Eastport Formation (31,76,77)
5 Hedgehog Formation (31)
6 Quimby Formation (51)
7 Quoddy Formation (31,76,77)
8 Tomhegan Formation
Kineo Rhyolite Member (32)
9 Traveler Rhyolite (32,169)
10 Thorofare Andesite (46)
11 Vinalhaven Rhyolite (46)

- Unconformity
- Dome
- Basin
- Upright antiform
- Upright synform
- Inclined/recumbent antiform
- Inclined/recumbent synform
- Normal or strike-slip fault — undifferentiated
- Normal fault, relative motion indicated by U (up) and D (down)
- Strike-slip fault, relative motion indicated by arrows
- Thrust or reverse fault; barbs on upper plate
- Listric fault, hachures on upper plate
- Undivided plutonic rocks